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CAMBRIAN STRATIGRAPHY OF NORTHWESTERN WYOMING[1]

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ABSTRACT

The Flathead and Gros Ventre formations of the Teton, Gros Ventre, northwestern Wind River, and western Owl Creek Mountains gradually merge toward the east, forming the lower part of the Wyoming "Deadwood"; the Gallatin formation maintains its lithologic identity eastward into the Bridger Mountains, where it forms the upper part of the Wyoming "Deadwood." The Gallatin contains a number of faunal zones corresponding to zones in the type St. Croixian, the lowest zone containing an early Upper Cambrian fauna. Paleontologic and stratigraphic evidence indicates that the pre-Gallatin formations are largely Middle Cambrian.

The formational name, *Depass*, is proposed for beds lying below the Gallatin in the Wyoming "Deadwood." A persistent limestone member in the lower Gros Ventre and the lower limestone member of the Gallatin are designated by the names, *Death Canyon* and *Du Noir*, respectively. The lithology, faunal content, and stratigraphic relationships of the formations are described, and an interpretation of the Cambrian history and paleogeography of Wyoming is presented.

INTRODUCTION

The Cambrian rocks of northwestern Wyoming, in and to the west of the Wind River Mountains, attain a thickness of about 1,000 feet and are divided into the Flathead, Gros Ventre, and Gallatin formations. The lowest formation, the Flathead, consists of sandstone and quartzite; the overlying Gros Ventre formation consists largely of shale and contains a prominent limestone member in the

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lower part; the highest formation, the Gallatin, is composed almost wholly of limestone but has a persistent shaly member in the middle part. The Cambrian of the Owl Creek, Bridger, and Big Horn ranges, which lie to the northeast of the Wind River Mountains, has

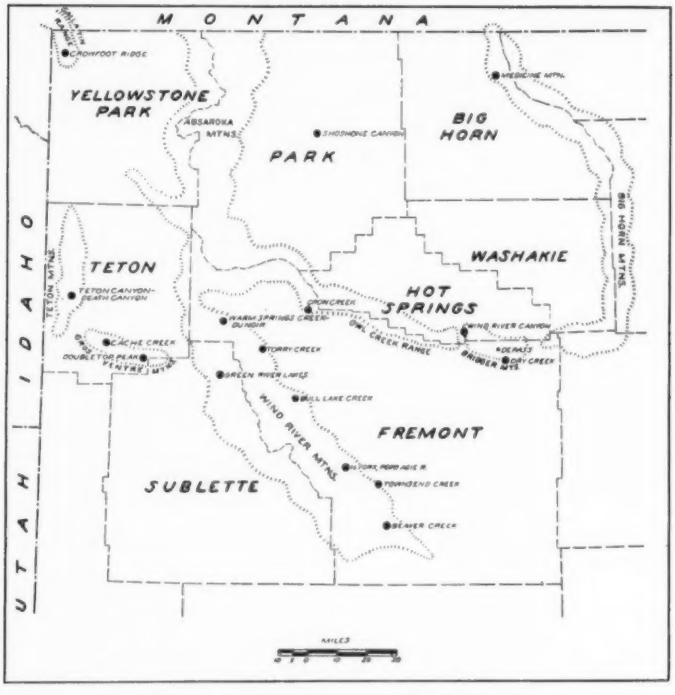

FIG. 1.—Index map of northwestern Wyoming

been described by N. H. Darton as averaging about 900 feet in thickness and consisting of a lower sandstone member, a middle shale member, and an upper limestone member. Darton referred these beds to the Deadwood formation.[2]

Although no systematic paleontologic study of these formations

[2] N. H. Darton, "Geology of the Big Horn Mountains," *U.S. Geol. Surv. Prof. Paper 51* (1906), pp. 23–26; "Geology of the Owl Creek Mountains," *U.S. 59th Cong. 1st Sess. Senate Doc. 219* (1906), pp. 14–15.

has been undertaken in the past, Cambrian fossils from various localities in northwestern Wyoming have been described by C. D. Walcott; and most authors, probably basing their conclusions largely on Walcott's work, have assigned the Flathead and Gros Ventre to the Middle Cambrian and the Gallatin to the Upper Cambrian. The most recent work indicates that the Deadwood, at its type section in the Black Hills, is of Upper Cambrian age.[3] The stratigraphic relationships and gradations in lithology between the Cambrian sections in the various mountain ranges of northwestern Wyoming have never been described, nor has there been any attempt to make accurate correlations between the Wyoming Cambrian and the standard Cambrian sections of other regions.

The field work here reported, which was carried out during the summers of 1933 and 1934, consisted primarily in tracing the Flathead, Gros Ventre, and Gallatin formations from the Teton Mountains through the Gros Ventre Mountains and northwestern part of the Wind River Mountains into the Wyoming Deadwood of the Owl Creek and Bridger ranges. In addition, detailed studies were carried out in the Wind River Mountains as far to the southeast as the Atlantic City district; and several days of reconnaissance work were spent in the Big Horn Mountains and in the Rattlesnake–Cedar Mountain uplift west of Cody.

EVOLUTION OF THE NOMENCLATURE

Flathead–Gros Ventre–Gallatin sequence.—In 1893, A. C. Peale divided the Cambrian of the Three Forks district, Montana, into two formations.[4] The lowest formation, the Flathead, was subdivided into the "Flathead quartzite" and "Flathead shales"; the Gallatin formation, overlying the Flathead, was subdivided into the "Trilobite limestones," "Obolella shales," "Mottled limestones," "Dry Creek shales," and "Pebbly limestones" (Fig. 2).

In 1899, Iddings and Weed also used the names "Flathead" and "Gallatin" in describing the Cambrian at Crowfoot Ridge in the

[3] Howard A. Meyerhoff and Christina Lochman, "*Crepicephalus* Horizon in the Deadwood Formation of South Dakota," *Geol. Soc. Amer. Proc. for 1933* (1934), p. 99.

[4] A. C. Peale, "The Paleozoic Section in the Vicinity of Three Forks, Montana," *U.S. Geol. Surv. Bull. 110* (1893), pp. 20–25.

Gallatin Mountains of Yellowstone Park.[5] However, they applied them in a different way, making the "Mottled limestone" the basal member of the Gallatin and expanding the use of the name "Flathead" to include all of the underlying Cambrian beds.

In 1918, Blackwelder restricted the name "Flathead," to the basal sandstones, and applied the name "Gros Ventre" to the shales and limestones intervening between the restricted Flathead and the Gallatin as redefined by Iddings and Weed.[6]

In 1900, W. H. Weed used the names "Flathead," "Wolsey," "Meagher," "Park," "Pilgrim," "Dry Creek," and "Yogo" for Peale's "Flathead quartzite," "Flathead shales," "Trilobite limestones," "Obolella shales," "Mottled limestones," "Dry Creek shales," and "Pebbly limestones," respectively.[7] At present Weed's names are being used in the Little Belt and Big Belt districts of Montana, although the names "Flathead," "Gros Ventre," and "Gallatin" are being used in the Three Forks district, as well as in northwestern Wyoming.[8]

In present usage, the name "Flathead" is applied to the basal sandstones and quartzites, even though they may vary in age from place to place; and that usage is continued in this paper. In Wyoming the name "Gallatin" is applied to the beds between the base of the massive, mottled, oölitic limestones and the Ordovician Big-Horn dolomite. The name "Gros Ventre" is applied to beds between the Flathead and Gallatin. The base of the Gros Ventre is drawn at the base of the first thick shales, and at places the Gros Ventre–Gallatin contact is disconformable.

So far as known to the writer, no one has systematically traced the Gallatin from the Three Forks district into Wyoming; and since the few fossils which have been described were not accurately listed as to

[5] J. P. Iddings and W. H. Weed, "Descriptive Geology of the Gallatin Mountains," *U.S. Geol. Surv. Mono. 32*, "Geology of Yellowstone Park," Part II (1899), p. 8.

[6] Eliot Blackwelder, "New Geological Formations in Western Wyoming," *Jour. Wash. Acad. Sci.*, Vol. VIII, No. 13 (1918), pp. 417–26.

[7] W. H. Weed, "Geology of the Little Belt Mountains, Montana," *U.S. Geol. Surv. 20th Ann. Rept.*, Part III (1900), pp. 284–87.

[8] M. Grace Wilmarth, *Tentative Correlation of the Named Geologic Units of Montana*, (U.S. Geol. Surv., July, 1932); *Tentative Correlation of the Named Geologic Units of Wyoming* (U.S. Geol. Surv., April, 1925).

their exact positions in the sections, there is no paleontologic proof that the base of the Gallatin is at precisely the same stratigraphic horizon from Montana to Wyoming (Fig. 2). In this paper the name

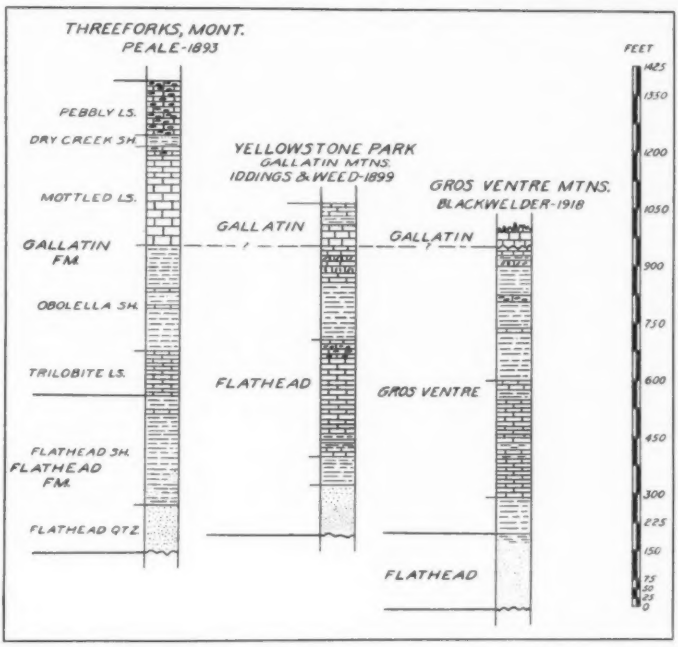

FIG. 2.—Type sections of the Flathead, Gros Ventre, and Gallatin formations. The columnar sections have been drafted by the writer from the text descriptions of the authors.

will be used in the same way as it has been used in the Wind River, Gros Ventre, and Teton Mountains.[9]

Wyoming Deadwood.—In 1901 Darton applied the name "Deadwood"[10] to the Cambrian rocks of the Black Hills and later applied

[9] D. Dale Condit, "Phosphate Deposits in the Wind River Mountains near Lander, Wyoming," *U.S. Geol. Surv. Bull. 764* (1924), pp. 7–8; Blackwelder, *op. cit.*; A. R. Schultz, "A Geologic Reconnaissance for Phosphate and Coal in Southeastern Idaho and Western Wyoming," *U.S. Geol. Surv. Bull. 680* (1918), pp. 17–19.

[10] N. H. Darton, "Description of the Geology and Water Resources of the Southern Half of the Black Hills and Adjoining Regions, South Dakota and Wyoming," *U.S. Geol. Surv., 21st Ann. Rept.*, Part IV (1901), pp. 505–8.

the same name in the Big Horn and Owl Creek Mountains.[11] The present writer does not apply "Deadwood" in the Owl Creek and Bridger Mountains, and questions whether it should be applied in the Big Horns.

STRATIGRAPHY OF THE FLATHEAD FORMATION

The Flathead formation averages about 175 feet in thickness and consists of quartz sandstones and quartzites which lie unconformably on the pre-Cambrian rocks, mostly granite, and conformably underlie the Gros Ventre shales. The basal beds of the formation as a rule are conglomeratic and arkosic; and in places the contact with the granite is more or less "blended," the Flathead arkose grading into weathered granite, which in turn gradually passes into unweathered granite. The upper beds of the formation usually contain thin layers of shale, and the boundary with the Gros Ventre is drawn at the base of the lowest thick shale series.

In sections studied by the writer the thickness of the Flathead varies from 225 to 60 feet. This variation is extremely irregular and does not appear to be consistent in any one direction. For example, at Du Noir, in the northwestern Wind River Mountains, the formation is 190 feet thick, whereas at Torry Creek, 15 miles to the southeast, it is only 60 feet thick, and in sections still farther to the southeast the thickness is comparable to that at Du Noir. These variations are believed by the writer to reflect the relief of the pre-Cambrian peneplain over which the Flathead sea advanced, the beds thinning by overlap against numerous elevations on the crystalline surface, rather than thickening because of local piling-up of sediment on the sea floor.

The formation is often referred to as the Flathead "quartzite," although in northwestern Wyoming the only quartzites seen are in the Teton and Gros Ventre Mountains, where they are restricted to about 85 feet in the middle part of the formation; in sections to the east of the Gros Ventres the Flathead is sandstone. The top of the Flathead commonly forms a terrace platform or dip slope projecting

[11] N. H. Darton, "Comparison of the Stratigraphy of the Black Hills, Big Horn Mountains, and Rocky Mountain Front Range," *Geol. Soc. Amer. Bull. 15* (1904), p. 395; "Geology of the Big Horn Mountains," *loc. cit.*; "Geology of the Owl Creek Mountains," *loc. cit.*

from beneath the shales of the Gros Ventre. The formation is in most cases poorly exposed.

The only fossil species found in the Flathead by the writer is *Lingulepis acuminatus*, which occurs in about 25 feet of beds at the top of the formation and ranges into the basal beds of the Gros Ventre.

STRATIGRAPHY OF THE GROS VENTRE FORMATION

The Gros Ventre formation, averaging about 675 feet in thickness, consists of shales and limestones which intervene between the Flathead and Gallatin formations. The lower shales of the Gros Ventre are arenaceous and contain thin beds of sandstone, and in places the boundary with the Flathead must be drawn arbitrarily. The top of the formation, in the Teton and Gros Ventre Mountains, is marked by a pronounced erosional contact with the Gallatin; in the northwestern Wind River Mountains evidence of erosion at the top of the Gros Ventre is not pronounced and is present only locally; and in the Owl Creek Range the Gros Ventre and Gallatin appear to be completely conformable.

Lower shale division.—The lowest division of the Gros Ventre consists largely of greenish-gray, soft, micaceous shales which average about 100 feet in thickness. The lower part of the shales contains layers of sandstone from a fraction of an inch to one foot thick; these sandstone beds are seen to increase in number and thickness as the formation is followed eastward. The upper part contains beds of calcareous sandstone and argillaceous limestone. The only fossils found in this division are *Lingulepis acuminatus*, which occurs in the sandstones in the lower part, and numerous, irregularly branching, tubular markings, which possibly represent worm borings. Complete exposures of this division are rare, the shales commonly weathering into a steep slope largely covered by talus from the limestones above.

Death Canyon member.[12]—The second division of the Gros Ventre is designated in this paper as the "Death Canyon" member, the type section being along the divide between Death and Teton canyons in the Teton Range. It consists largely of fine-grained, dark-gray and black limestone mottled with brown. The limestones are thin-

[12] "Death Canyon," new member name (type section, Table I).

bedded for the most part, occurring in layers from one-half to several inches thick, separated by argillaceous seams and partings. Trilobite fragments are present occasionally along the partings, although for the most part the specimens are too poorly preserved for identification. In the Teton Mountains, along the divide between Death and Teton canyons, a faunule consisting of *Glyphaspis* sp., *Bolaspis* (?) *resseri*, and *Marjumia* (?) *tetonensis* was found on talus blocks about 100 feet below the top of the member; in the Gros Ventre Mountains, about 2 miles east of Cache Creek, the same faunule occurs in place 75 feet below the top. Aff. *Kochaspis* sp. occurs in the upper 20 feet of the Death Canyon member at Crow Creek Canyon in the western Owl Creeks. "Worm borings" also occur, but are not so abundant as in the shale divisions.

The Death Canyon member attains a thickness of 285 feet in the Teton Range and gradually thins and interfingers with shale toward the east, disappearing in the middle parts of the Owl Creek and Wind River ranges. At Du Noir, in the northwestern Wind River Mountains, the Death Canyon is 219 feet thick and contains numerous shale beds in the lower part; at Torry Creek it is about 100 feet thick; at Bull Lake Creek only 40 feet of it are exposed, and it is probably not much thicker than the outcrop; along the North Fork of the Popo Agie River there is no evidence of the Death Canyon, nor was it seen in any sections farther to the southeast. In the Owl Creek Mountains the Cambrian is usually very poorly exposed, and it is not possible to trace the eastward thinning of the Death Canyon as easily as in the Wind Rivers. The greater part of the Death Canyon beds commonly outcrop in a prominent scarp in the Teton and Gros Ventre Mountains, the scarp becoming less prominent in the Wind Rivers and Owl Creeks because of thinning and interfingering of the limestones with shales.

Upper shale division.—The upper division of the Gros Ventre varies considerably in thickness, which is in the order of 200 to 300 feet. It consists largely of greenish-gray shales similar to the shales of the lower division. In places the shales are purplish and dark-green, owing to the presence of hematite and glauconite.

Limestone beds varying from an inch to several feet in thickness are common in the upper shale division. Some of the limestones are

dense, light-gray, brown, or greenish-gray, and are either laminated with micaceous seams or consist of thin layers and lenses separated by shaly partings, whereas others are of the type generally referred to as "intraformational conglomerate" (Fig. 3). In this paper these conglomerates are divided into "intraformational breccias" and "pebbly limestones." The breccias consist of thin limestone plates and lenses, of irregular outline and averaging 6 inches to 1 or 2 feet in diameter, which lie at all angles to the bedding, whereas the pebbly

FIG. 3.—Intraformational breccia in the upper part of the Gallatin formation in Shoshone Canyon.

limestones consist of small, flat-to-rounded pebbles which mostly lie parallel or nearly parallel to the bedding. In both types the pebbles or plates are cemented in a matrix which is usually coarse-grained, light-gray, glauconitic limestone, although in some cases the matrix is fine, argillaceous limestone. Beds consisting of thin layers and lenses of limestone with shaly partings are invariably associated with the breccias and in places can be seen to pass upward and laterally into thick breccia beds. In some cases the limestone plates in the breccias appear to be more or less bent; and since they show no evidence of having been rounded or washed by waves, it is considered possible that the breccias are autoclastic rocks produced by the slipping or slumping of the thin-layered limestones along their argil-

laceous partings while the sediment was still in a semiplastic condition. However, the pebbles of the pebbly limestones are more or less rounded and probably originated through waves breaking up beds of platy limestone or breccia.

The upper 50–100 feet of the upper division of the Gros Ventre, in the Owl Creek and Wind River Mountains, is composed largely of thin-bedded limestones, pebbly limestones, and intraformational breccias. These beds are absent for the most part in the Teton and

FIG. 4.—Disconformable contact (at hammer head) of the Gros Ventre and Gallatin in the South Fork of Teton Canyon. The beds on the lower right are thin-bedded, lenticular limestones which grade vertically and laterally into intraformational breccias lying above and to the left.

Gros Ventre Mountains, where the Gros Ventre is about 100 feet thinner than in sections to the east (Fig. 5); the missing limestones probably are represented by the pronounced disconformity at the top of the Gros Ventre. In the northwestern Wind River and western Owl Creek Mountains, the intraformational breccias always attain their maximum development just below the top of the Gros Ventre, where they form individual beds 4–5 feet thick.

Fragmentary trilobites are common in the limestones of the upper division of the Gros Ventre, as well as numerous "worm borings." At Crow Creek Canyon a fauna consisting of *Marjumia* (?) *tipperaryensis*, *Crepicephalus* (?) *wyomingensis*, several species of *Brachyaspis*, and *Dicellomus* sp. occur in lenticular limestones about 120

feet above the top of the Death Canyon member. In the Wind River and Owl Creek Mountains, oboloid brachiopods characterize the limestones in the upper part of this division.

The shales of the upper division are usually poorly exposed,

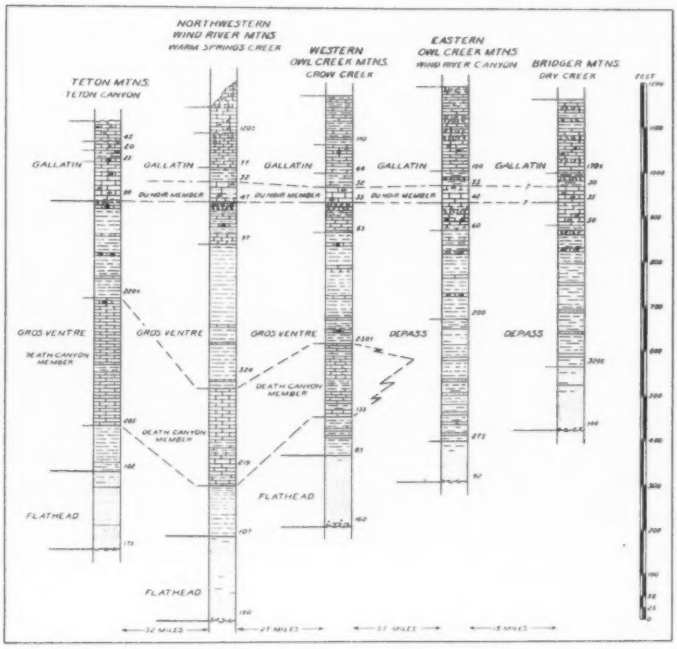

FIG. 5.—Correlated sections of the Cambrian formations from the Teton Range to the Bridger Range.

weathering back to form a steep slope. In the Wind River Mountains and in sections to the east, the limestones at the top of the Gros Ventre form a scarp in association with the massive limestones of the lower Gallatin.

STRATIGRAPHY OF THE DEPASS FORMATION[13]

As the Flathead and Gros Ventre formations are followed eastward through the Owl Creek and Wind River Mountains, the Flathead sandstone lithology is seen to rise gradually and invade the

[13] "Depass," new formational name (type section, Table IV).

lower part of the Gros Ventre, the sandstone and shale facies interfingering and finally becoming so greatly mixed that the two formations are not separable and are designated together as the "Depass" formation (Fig. 5). The type section is in the Wind River Canyon at the eastern end of the Owl Creek Range, where the Depass is defined as all the beds in the Wyoming "Deadwood" below the Gallatin. The name is taken from the "D" Pass (now spelled "Depass"), near the eastern end of the Bridger Range.

The name "Depass" will apply from about the middle of the Owl Creek Range at least as far to the east as the eastern end of the Bridger Range, and it is probable that it will apply in part of the Big Horns. In the Wind River Mountains it will apply from the southeastern end of the range at least as far to the northwest as the North Fork of the Popo Agie River, but not farther north than Bull Lake Creek. The Cambrian nomenclature to the southeast of the Wind River Mountains, in central Wyoming, must be considered as undetermined.

Well-exposed sections of the Depass formation are not common. The sandstones in the lower part generally outcrop as low ledges, the shales in the middle part form long slopes, and the limestones in the upper part usually form a scarp in association with the lower limestones of the Gallatin. The description of the type section (Table IV) and of the section at Dry Creek (Table V) present the character of the formation.

STRATIGRAPHY OF THE GALLATIN FORMATION

The Gallatin formation, in the area studied by the writer, attains its maximum thickness of about 300 feet in the northwestern part of the Wind River Mountains, thinning somewhat toward the southeastern end of the range. In the Gros Ventre and Teton Mountains it is about 180 feet thick; its average thickness in the upper part of the Wyoming "Deadwood" of the Owl Creek–Bridger uplift is about 225 feet. Although the Gallatin is somewhat variable in character, for purposes of description it can be divided into a lower limestone division, a middle shaly division, and an upper limestone division.

Du Noir member.[14]—The lower limestone division of the Gallatin,

[14] "Du Noir," a new member name (type section, Table II).

as it occurs in the Wind River Mountains and Owl Creek–Bridger uplift, is designated in this paper as the "Du Noir" member. The

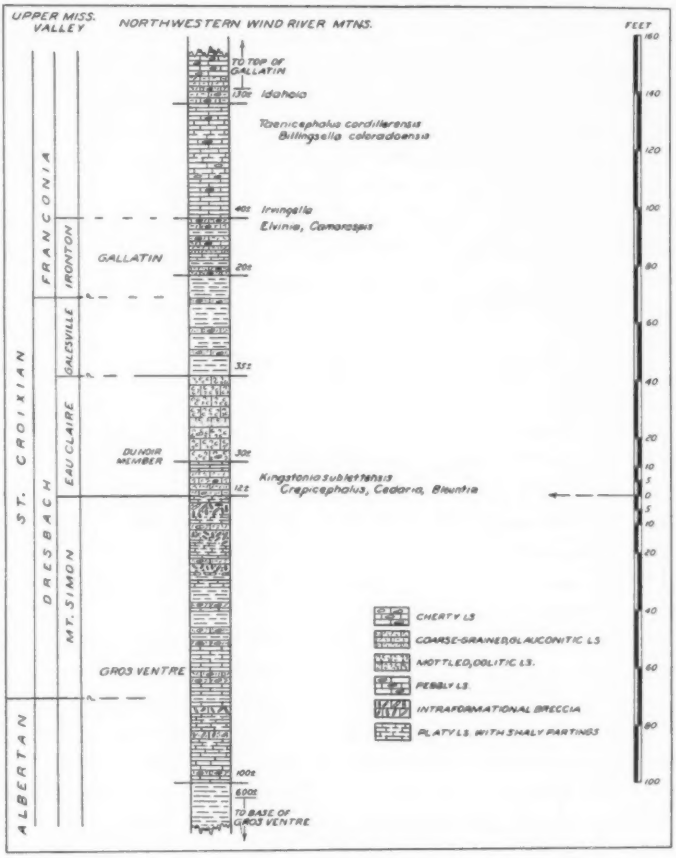

FIG. 6.—Generalized section of the upper Gros Ventre and lower Gallatin. The column on the left shows the correlation with the type St. Croixian.

type section is along Warm Springs Creek, 2 miles west of Du Noir, in the northwestern part of the Wind River Mountains. The Du Noir is composed largely of massive, dark-gray limestone mottled with yellowish-brown and filled with brown oölites. In most sections

it contains 1–3 feet of calcareous sandstone in the middle part and usually has about 10 feet of laminated limestone and glauconitic, pebbly limestone at the base. The average thickness is about 40 feet.

Trilobites are rather common in the Du Noir, although for the most part they are too fragmentary for identification. A fauna characterized by *Kingstonia sublettensis* and several species of *Crepicephalus* occurs in the basal glauconitic limestones at a number of localities in the Wind River Mountains; *Blountia* and *Cedaria* are other significant genera associated with this fauna.

In the Gros Ventre and Teton Mountains the Du Noir member is represented in the lower part of the lower limestone division of the Gallatin. However, the lower limestones are about twice as thick as in the Wind River Mountains, and the name "Du Noir" is not applied (Fig. 5).

The lower limestones of the Gallatin are massive and resistant for the most part, forming a prominent scarp.

Middle shaly member.—In the Wind River Mountains and Owl Creek–Bridger uplift, the Gallatin, for an interval of 30–35 feet above the Du Noir member, is composed largely of greenish-gray shale with beds of pebbly, glauconitic limestone and nodular, bluish-green, argillaceous limestone. Good exposures are rare, the beds generally weathering into a slope covered by talus from the limestone scarp above.

The middle shaly division in the Gros Ventre and Teton Mountains is 20–25 feet thick and does not contain as much shale as does the corresponding member in the Wind River Mountains. If this shaly facies is stratigraphically continuous from the Wind Rivers to the Gros Ventres and Tetons, it has risen 20 or 30 feet in the Gallatin, inasmuch as it contains the *Elvinia* fauna in the latter ranges; whereas in the Wind River Mountains that fauna characterizes the lower part of the limestone division overlying the middle shaly beds.

Upper limestone division.—The upper limestone division of the Gallatin attains its maximum thickness of about 195 feet in the northwestern part of the Wind River Range, thinning somewhat in the Owl Creek and Bridger ranges. To the west of the Wind River Range, in the Gros Ventre and Teton ranges, the thickness is only about 70 feet; this difference is due in part to the higher stratigraphic

position of the middle shaly member, as previously explained, and in part to erosion, the missing beds being represented by disconformities within, and at the top of, the Gallatin (Fig. 5).

The upper limestone division of the Gallatin contains a sequence of distinct faunal zones (Fig. 6). These zones appear to be more persistent in the Wind River Mountains than elsewhere. There the lower 20 feet are composed largely of glauconitic, pebbly limestones characterized by *Elvinia* and *Camaraspis*. Overlying the *Elvinia* beds are about 40 feet of dove-gray, thin-bedded limestones characterized by *Irvingella* in the lower part and by *Taenicephalus*, *Wilbernia*, and *Billingsella coloradoensis* in the upper part. The beds immediately overlying the *Taenicephalus* zone, for an interval of about 15 feet, are mostly coarse-grained, pebbly limestones characterized by *Idahoia*.

The remaining beds of the Gallatin above the *Idahoia* zone reach their maximum thickness in the northwestern part of the Wind River Range, where they are composed of gray, tan, and cream-colored limestones; their equivalents in the Owl Creek–Bridger uplift contain beds of shale and much pebbly limestone and intraformational breccia. Although no persistent faunal zones have been determined in these beds, several brachiopod and trilobite genera, including *Briscoia*, occur in the upper part.

CAMBRIAN SECTIONS IN NORTHWESTERN WYOMING

TETON MOUNTAINS

A well-exposed Cambrian section occurs along the divide between the South Fork of Teton Canyon and Death Canyon, about 2 miles south of Buck Mountain. The Cambrian is considerably thinner than in the Wind Rivers, the missing beds being represented largely by disconformities at the top of the Gros Ventre and at the top of the Gallatin. The type occurrence of the Death Canyon member is in this section, the limestones forming a prominent, double-rimmed scarp (Fig. 7). The massive lower limestones of the Gallatin are about twice as thick as in the Wind Rivers, and only in places do the beds exhibit the mottled appearance and brown oölites characteristic of the Du Noir member; beds equivalent to the *Kingstonia-Crepicephalus* zone are presumably present at the base of the lower Galla-

tin limestones, although the fauna has not been found. The *Elvinia* and *Irvingella* zones have been located and occur at about the same distance above the base of the Gallatin as in the Wind Rivers. The *Taenicephalus* and *Idahoia* zones have not been located and may be missing, inasmuch as there is a distinct disconformity within the Gallatin just above the *Irvingella* zone.

FIG. 7.—View in the South Fork of Teton Canyon. The double-rimmed scarp is the Death Canyon member of the Gros Ventre; the blocks in the foreground are at the top of the Flathead.

TABLE I

SECTION ALONG THE DIVIDE BETWEEN THE SOUTH FORK OF TETON CANYON AND DEATH CANYON, TETON COUNTY, WYOMING

Ordovician Big Horn formation: dolomite	
- - - - - - - - - - - - - - - - - Erosional disconformity - - - - - - - - - - - - - - - - -	
Cambrian	962 ft. (total)
Gallatin formation	180 ft. (total)
At base 3 ft. dove-gray limestone grading upward into light-gray to white limestone weathering to granular tan surface. "Worm-borings" and chert nodules common. Lower 20 ft. contain cystid (?) cups and abundant *Billingsella coloradoensis*	45 ft.
- - - - - - - - - - - - - - - - - Erosional disconformity - - - - - - - - - - - - - - - - -	
Dove-gray, fine-grained, thin-bedded limestone with "worm-borings." Top 5 ft.: *Irvingella gibba;* lower 15 ft.: *Eoorthis* (?) sp. cf. *E. iddingsi, Linarsonella* sp.	20 ft.

TABLE I—*Continued*

Largely covered. Gray, coarse, pebbly limestone with beds of soft, greenish-gray shale. *Elvinia roemeri*, *Pterocephalia* sp. cf. *P. sancti-sabae*, *Dunderbergia* (?) sp. cf. *D.* (?) *declivita*	25 ft.
Massive limestone. Mainly gray, mottled with greenish-yellow or brown; top 5 ft. yellow and yellowish-gray. In places pebbly, elsewhere with brown oölites. Trilobite fragments	90 ft.
Erosional disconformity	
Gros Ventre formation	607 ft. (total)
Largely covered. Greenish-gray shale with beds of pebbly limestone and intraformational breccia. Top 17 ft. is gray, thin-bedded limestone, pebbly limestone, and intraformational breccia	220 ft.
Death Canyon member (type)	285 ft. (total)
Black and dark-gray limestone mottled with brown; fine-grained, thin-bedded, with argillaceous seams and partings. Top 5 ft. is massive limestone weathering in spherical masses 3–5 ft. through. Top 15 ft.: *Glyphaspis* sp. cf. *G. perconcava;* talus slabs 100 ft. from top: *Marjumia* (?) *tetonensis Bolaspis* (?) *resseri*, *Glyphaspis* sp. und.	150 ft.
Greenish-gray shale and limestone similar to above	15 ft.
Limestone similar to above	112 ft.
Light-gray, fine-grained, thin-bedded limestone weathering light brown	8 ft.
Largely covered. Mainly greenish-gray, soft shale	80 ft.
Purplish and greenish, hard, arenaceous shales with a few thin beds of argillaceous sandstone containing *Lingulepis acuminatus*	22 ft.
Flathead formation	175 ft. (total)
Reddish and brownish, fine to medium-textured sandstone with beds of green, arenaceous shale	35 ft.
White and gray quartzite in beds up to 15 ft. thick	85 ft.
Red, purple, and gray, coarse, arkosic sandstone. Cross laminated in places	55 ft.
Unconformity	
Pre-Cambrian: Pink and gray granite containing xenoliths of gneiss and schist and cut by pegmatites and basic dikes.	

GROS VENTRE MOUNTAINS

The Cambrian is well exposed in the Gros Ventre Mountains about a mile east of Cache Creek and 9 miles southeast of Jackson. The section studied is located where the trail from Flat Creek to

Cache Creek crosses the divide between the two streams. The sequence is practically the same as at Teton and Death canyons.

Blackwelder has described the section at Doubletop Peak, about 18 miles southeast of the Cache Creek section.[15]

FIG. 8.—Section along Warm Springs Creek west of Du Noir. The blocks in the foreground are Flathead; the prominent lower cliff is the upper part of the Death Canyon. The Gros Ventre–Gallatin contact is just above the base of the second cliff above the Death Canyon; the Gallatin–Big Horn contact is in the middle part of the cliff capping the hill.

NORTHWESTERN WIND RIVER MOUNTAINS

The best Cambrian section in the northwestern part of the Wind River Range occurs along Warm Springs Creek. Most of the Cambrian is well exposed along the north side of the creek about 2 miles west of Du Noir, although there are fair exposures on Warm Spring

[15] *Loc. cit.*

Mountain about a mile east of Du Noir. At Warm Spring Mountain there are about 100 feet of limestones at the top of the Gallatin which are absent in the section west of Du Noir and apparently were eroded out prior to the deposition of the Big Horn.

TABLE II

SECTION ALONG WARM SPRINGS CREEK, NEAR DU NOIR, FREMONT COUNTY, WYOMING

Ordovician Big Horn formation: dolomite	
---------- Erosional disconformity ----------	
Cambrian	1,213 ft. (total)
Gallatin formation	276 ft. (total)
Light-gray and dark-gray, fine- to medium-grained limestone in layers a fraction to several inches thick. Chert nodules and "worm-borings" common. Contains a few beds of pebbly limestone. Top 5 ft.: *Briscoia schucherti;* at 40 ft. above base: *Synthrophia alata, Eoorthis* sp. cf. *E. iophon*	62 ft.
Partly covered. Mainly yellowish-gray to white, medium-grained limestone in layers from an inch to a foot thick. "Worm-borings" and cherty lenses common	58 ft.
---------- Local disconformity ----------	
Dove-gray, fine-grained, thin-bedded limestone marked by "worm-borings." Thin shaly layers common; in places pebbly. At 30–40 ft. above base: *Taenicephalus cordillerensis, Wilbernia* sp., *Billingsella coloradoensis, Billingsella* (?) sp. aff. *B. striata, Dicellomus* sp. und., *Lingulepis acuminatus* var. *meeki, Acrotreta microscopica* var. *tetonensis;* basal 5 ft.: *Irvingella gibba*	57 ft.
Gray limestone with shale partings. Some beds fine-grained, thin-layered; others coarse, glauconitic, pebbly. Top 5 ft.: *Elvinia roemeri, Burnetia* (?) sp. und., *Eoorthis* (?) sp. cf. *E. iddingsi*	20 ft.
Largely covered. Soft, greenish-gray shale with beds of bluish-green, soft, argillaceous limestone and pebbly limestone	32 ft.
Du Noir member (type)	47 ft. (total)
At base 2 ft. gray and yellow, fine-grained, calcareous sandstone grading upward into massive, dark-gray limestone mottled with yellow and filled with brown oölites. Trilobite fragments	18 ft.
Limestone similar to foregoing	16½ ft.

Cache Creek crosses the divide between the two streams. The sequence is practically the same as at Teton and Death canyons.

Blackwelder has described the section at Doubletop Peak, about 18 miles southeast of the Cache Creek section.[15]

FIG. 8.—Section along Warm Springs Creek west of Du Noir. The blocks in the foreground are Flathead; the prominent lower cliff is the upper part of the Death Canyon. The Gros Ventre–Gallatin contact is just above the base of the second cliff above the Death Canyon; the Gallatin–Big Horn contact is in the middle part of the cliff capping the hill.

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The best Cambrian section in the northwestern part of the Wind River Range occurs along Warm Springs Creek. Most of the Cambrian is well exposed along the north side of the creek about 2 miles west of Du Noir, although there are fair exposures on Warm Spring

[15] *Loc. cit.*

Mountain about a mile east of Du Noir. At Warm Spring Mountain there are about 100 feet of limestones at the top of the Gallatin which are absent in the section west of Du Noir and apparently were eroded out prior to the deposition of the Big Horn.

TABLE II

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Ordovician Big Horn formation: dolomite	
---------- Erosional disconformity ----------	
Cambrian	1,213 ft. (total)
Gallatin formation	276 ft. (total)
Light-gray and dark-gray, fine- to medium-grained limestone in layers a fraction to several inches thick. Chert nodules and "worm-borings" common. Contains a few beds of pebbly limestone. Top 5 ft.: *Briscoia schucherti;* at 40 ft. above base: *Synthrophia alata*, *Eoorthis* sp. cf. *E. iophon*	62 ft.
Partly covered. Mainly yellowish-gray to white, medium-grained limestone in layers from an inch to a foot thick. "Worm-borings" and cherty lenses common	58 ft.
---------- Local disconformity ----------	
Dove-gray, fine-grained, thin-bedded limestone marked by "worm-borings." Thin shaly layers common; in places pebbly. At 30–40 ft. above base: *Taenicephalus cordillerensis*, *Wilbernia* sp., *Billingsella coloradoensis*, *Billingsella* (?) sp. aff. *B. stricta*, *Dicellomus* sp. und., *Lingulepis acuminatus* var. *meeki*, *Acrotreta microscopica* var. *tetonensis;* basal 5 ft.: *Irvingella gibba*	57 ft.
Gray limestone with shale partings. Some beds fine-grained, thin-layered; others coarse, glauconitic, pebbly. Top 5 ft.: *Elvinia roemeri*, *Burnetia* (?) sp. und., *Eoorthis* (?) sp. cf. *E. iddingsi*	20 ft.
Largely covered. Soft, greenish-gray shale with beds of bluish-green, soft, argillaceous limestone and pebbly limestone	32 ft.
Du Noir member (type)	47 ft. (total)
At base 2 ft. gray and yellow, fine-grained, calcareous sandstone grading upward into massive, dark-gray limestone mottled with yellow and filled with brown oölites. Trilobite fragments	18 ft.
Limestone similar to foregoing	16½ ft.

TABLE II—*Continued*

Brown, gray, and greenish, coarse-grained, pebbly limestone sprinkled with black and green glauconite grains. *Kingstonia sublettensis*, *Blountia globosa*, *B. dunoirensis*, *B.* sp. cf. *B. amage*, *Crepicephalus* sp., *Cedaria* sp. aff. *C. prolifica*, *Arapahoia levis*, *Maryvillia* sp., *Agnostus* sp. cf. *A. tumidosus*	12½ ft.
Gros Ventre formation	747 ft. (total)
Gray, fine-grained, laminated, and thin-bedded limestones with shaly partings; pebbly limestones and much intraformational breccia, the last-mentioned forming beds 5 ft. in thickness near the top. Upper 10 ft.: *Agnostus* sp., *Linnarsonella* sp. cf. *L. modesta*, *L. transversa*, *Obolus* (?) sp. cf. *O. zetus*	40 ft.
Largely covered. Mainly gray, laminated limestone and glauconitic pebbly limestone with beds of greenish-gray shale. Trilobite fragments	57 ft.
Largely covered. Mainly soft, greenish-gray, micaceous shale with beds of laminated limestone and pebbly limestone up to 4 ft. thick. "Worm-borings" common	324 ft.
Death Canyon member	219 ft. (total)
Thin-bedded, dark-gray limestone mottled with brown. Trilobite fragments	115 ft.
Greenish-gray shale and limestone similar to above	20 ft.
Gray and pinkish limestone weathering rusty brown. Shale partings	46 ft.
Mainly gray and brown, arenaceous and micaceous limestone with thin shale beds	38 ft.
Greenish-gray and yellowish, micaceous shale with layers of gray and brown sandstone up to 10 in. thick	107 ft.
Flathead formation	
White, gray, yellow, brown, red sandstones in beds up to 4 ft. thick. Conglomeratic and arkosic at base; in places rippled-marked, cross-laminated. Upper 20 ft.: *Lingulepis acuminatus*	190 ft. (total)
Unconformity	
Pre-Cambrian: Pink and gray granite cut by pegmatites	

OWL CREEK–BRIDGER UPLIFT

Crow Creek Canyon.—The section next described is located in Crow Creek Canyon about 3 miles north of Tipperary Post-Office, Fremont County, near the western end of the Owl Creek Range.

The Cambrian and Ordovician make up a fault block which is downdropped along the eastern side of the Canyon, the beds dipping northeast and abutting against the pre-Cambrian crystallines which almost entirely surround the block. The beds are largely covered by talus, and the section is difficult to measure because of rapid changes in dip and strike.

FIG. 9.—Crow Creek Canyon section. The triple-rimmed ledge in the middle foreground is the Death Canyon; the first cliff above the Death Canyon is the Du Noir member of the Gallatin.

TABLE III

SECTION IN CROW CREEK CANYON, FREMONT COUNTY, WYOMING

Ordovician Big Horn formation: At base 3 ft. of yellowish and pinkish quartzite grading upward into massive dolomite. Exact contact with Gallatin not exposed.	
Cambrian	965 ft. (total)
Gallatin formation	241 ft. (total)
Largely covered. Greenish-gray shale and limestones of various types: gray limestone mottled yellow and brown, fine-grained, in beds 1–3 ft. thick; gray and pinkish, fine-grained, thin-bedded limestone; glauconitic pebbly limestone. Some layers with trilobite fragments	110 ft.
Mainly dove-gray, fine-grained, thin-bedded limestone with some light-gray, glauconitic, pebbly limestone. Thin shale layers common throughout. Upper 15 ft.: *Idahoia* sp. cf. *I. serapio*, *I.* sp. und.; at 15–25 ft. above base:	

TABLE III—*Continued*

Taenicephalus cordillerensis, Billingsella coloradoensis, B. sp. cf. *B. plicatella*	36 ft.
Light-gray, pebbly limestone and dark-gray, fine-grained limestone splashed with brown. *Eoorthis* (?) sp. cf. E. *iddingsi*	28 ft.
Largely covered. Mainly greenish-gray shale	32 ft.
Du Noir member	
Mottled dark-gray and brown, massive limestone filled with brown oölites	35 ft. (total)
Contact apparently conformable	
Gros Ventre formation	555± ft. (total)
Greenish-gray, fine-grained, laminated limestones, pebbly limestones, and intraformational breccia. The breccias form beds 3–4 ft. thick in the upper part. *Obolus* (?) sp. cf. *Ozetus*	65 ft.
Largely covered. Mainly greenish-gray, purplish, and dark-green shale with beds of gray, fine-grained limestone and pebbly limestone. At about 120 ft. above the base *Brachyaspis rhynchina, B.* (?) sp., *Crepicephalus* (?) *wyomingensis, Marjumia* (?) *tipperaryensis, Dicellomus* sp. (Note: stated thickness approximate, possibly 25–50 ft. less than it should be)	250± ft.
Death Canyon member	155 ft. (total)
Dark-gray, thin-bedded limestone mottled yellow and brown. Green and red argillaceous partings. Lower 10 ft.: aff. *Kochaspis* sp.	20 ft.
Red and green shale with beds of limestone similar to foregoing and pebbly limestone	11 ft.
Dark-gray, thin-bedded limestone mottled yellow and brown	48 ft.
Largely covered. Greenish-gray shale and argillaceous limestone	28 ft.
Dark-gray limestone mottled with brown	24 ft.
Greenish-gray shale	12 ft.
At base gray, calcareous sandstone grading upward into gray and yellowish, argillaceous limestone	12 ft.
Greenish-gray shale with beds of gray, red, and yellow sandstone containing *Lingulepis acuminatus*	85 ft.
Flathead formation	
White, gray, yellow, brown, red sandstones. Conglomeratic and arkosic at base; in places cross-laminated and ripple-marked	160 ft. (total)
---------- Unconformity ----------	
Pre-Cambrian: Pink and gray granite.	

Wind River Canyon.—The Cambrian is exceptionally well exposed along both sides of Wind River Canyon, which is usually considered to be the boundary marking the eastern end of the Owl Creek Range

FIG. 10.—Section above Smith's cabin, Wind River Canyon. The cliff in the foreground is of lower Depass sandstone. The Big Horn–Gallatin contact is just above the base of the massive limestone cliff forming the canyon rim. The Gallatin-Depass contact is at the base of the dark limestones in the upper part of the first cliff below the Big Horn.

and the western end of the Bridger Range. Most of the section described below, from the base of the Depass to the top of the Du Noir member of the Gallatin, was measured on the east side of the Canyon, just above Smith's cabin; the beds from the top of the Du Noir to the base of the Big Horn were measured on the east side of the canyon about one-quarter mile east of Boysen Dam.

TABLE IV

SECTION IN WIND RIVER CANYON,
FREMONT COUNTY, WYOMING

Ordovician Big Horn formation: At base 2 ft. of yellow sandstone grading upward into massive dolomite.

Contact apparently conformable

Cambrian	892 ft. (total)
Gallatin formation	265 ft. (total)
Red and pink, thin-bedded limestone and light-gray argillaceous limestone. Star-shaped echinoderm stems common in upper part	30 ft.
Mainly pebbly limestone and intraformational breccia. Purplish, thin-bedded argillaceous limestone prevails in the upper part	60 ft.
Mainly dove-gray, thin-bedded limestone. Contains several thick beds of intraformational breccia in the upper part. Talus slabs in lower part: *Taenicephalus cordillerensis*, *Saratogia* sp., *Billingsella coloradoensis*	100 ft.
Greenish-gray shale and intraformational breccia	33 ft.
Du Noir member	
Mainly massive limestone; some thin-bedded and some pebbly limestone. Contains several beds, 1–4 feet thick, of dark-gray limestone mottled with brown and filled with brown oölites. At 13–16 ft. below the top the beds vary from a light-gray, thin-bedded limestone to a gray calcareous sandstone. At 10–15 ft. below top: *Crepicephalus tripunctatus*, *Maryvillia* sp. cf. *M. ariston*, *Hyolithes* sp.	42 ft. (total)

Contact apparently conformable

Depass formation (type)	627 ft. (total)
Mainly gray, fine-grained, thin-bedded limestone with shaly partings. Contains some pebbly limestone and intraformational breccia. Lower 10 ft.: *Arapahoia spatulata*, *A.* sp. cf. *A. typa*, *Dicellomus nana*, *Linnarsonella tennesseensis*, *L.* sp. cf. *L. modesta*	60 ft.
Largely covered. Mainly greenish-gray shale containing beds of gray, laminated limestone, pebbly limestone, and intraformational breccia	200 ft.
Dark-green and purple, hard, arenaceous sandstone containing a few layers of glauconitic limestone. Brachiopod fragments	10 ft.

TABLE IV—*Continued*

Gray and yellowish sandstone and green, hard, arenaceous shale. The sandstones carry fragments of linguloid brachiopods	34 ft.
Mainly dark-green and purplish shale with thin layers of gray, fine-grained sandstone carrying brachiopod fragments	136 ft.
Green shale with beds of light-gray sandstone from a fraction of an inch to 3 ft. in thickness. Cross-laminated in places; "worm-borings" common. The sandstones carry fragments of linguloid brachiopods	95 ft.
Green shale and sandstone alternating in beds about 6 in. thick. Cross-laminated in places; "worm-borings" common	25 ft.
Greenish and yellowish, massive sandstones marked by vertical, tubular structures resembling *Scolithus*	18 ft.
Gray and yellowish sandstone with a few thin layers of green and reddish shale. Cross-lamination common throughout. Conglomeratic and arkosic at base where the beds grade downward through decomposed granite into unweathered granite	49 ft.
---------- Unconformity ----------	
Pre-Cambrian: Pink and gray granite and pegmatites intruding black phyllites and slates.	

Dry Creek.—The section described in Table V is located near the eastern end of the Bridger Range, in the northwest quarter of T. 39 N., R. 92 W. The Cambrian forms part of an "island" of Paleozoic rocks which rises above the Eocene sediments along Dry Creek, a tributary of Badwater Creek. The beds are poorly exposed and difficult to measure because of faulting and rapid changes in dip.

Because of poor exposures, it has not been possible to determine accurately the boundaries of the Du Noir member of the Gallatin; and inasmuch as the writer did not observe any beds exhibiting the strongly mottled appearance and brown oölites which characterize the member in sections to the west, the name has not been applied (Fig. 5). However, the Du Noir is easily recognized in Wind River Canyon, 15 miles to the west; and it is probable that the name could be applied at Dry Creek if the section were well exposed.

TABLE V

SECTION ALONG DRY CREEK, FREMONT COUNTY, WYOMING

Ordovician Big Horn formation: At base 3–7 ft. of dense, white quartzite followed by massive dolomite.	
Contact apparently conformable	
Cambrian	745± ft. (total)
Gallatin formation	235± ft. (total)
Mainly gray intraformational breccia with beds of shale. Upper 30 ft. is reddish, thin-bedded limestone and intraformational breccia	125± ft.
Dove-gray, thin-bedded limestones and intraformational breccia. Some talus slabs from this division carry *Wilbernia* (?) sp.; other slabs carry *Eoorthis remnicha*	45 ft.
Greenish-gray and purplish shale with beds of intraformational breccia and greenish-gray limestone. Upper 10 ft.: *Kyphocephalus bridgerensis*, *Dunderbergia* (?) *declivita*	30 ft.
Gray limestone and intraformational breccia. Beds in general more massive than those below	35± ft.
Exact contact not determinable	
Depass formation	510± ft. (total)
Mainly gray, thin-bedded, argillaceous limestone. Talus slabs from the lower part contain *Arapahoia spatulata* and *Dicellomus nana*	50 ft.
Greenish-gray shale with a few beds of limestone, mainly intraformational breccia	85 ft.
Dark-green and purplish, hard arenaceous shale with beds of sandstone from a fraction of an inch to several feet thick. The sandstones contain *Lingulepis acuminatus*	235± ft.
Gray sandstone with thin beds of arenaceous shale. Vertical, tubular markings resembling *Scolithus* are common in the sandstones	40 ft.
Pink, yellow, and gray sandstone. Arkosic and conglomeratic at base; some beds cross-laminated	100± ft.
Contact covered	
Pre-Cambrian: Pink and gray granite containing xenoliths of gray and black schist and cut by pegmatites and basic dikes.	

BIG HORN MOUNTAINS

The Cambrian of the Big Horn Mountains has been described by Darton and referred by him to the Deadwood formation.[16] However,

[16] N. H. Darton, "Geology of the Big Horn Mountains," *loc. cit.*

the Wyoming "Deadwood," as pointed out elsewhere in this paper (p. 141), evidently includes beds considerably older than the type Deadwood in the Black Hills; and since the writer has not applied the name, "Deadwood," in the Owl Creek–Bridger uplift, it is questionable whether it should be applied in the Big Horns.

Although the writer spent several days of reconnaissance study in the Big Horn Range, the work did not progress to the point where conclusions relative to the Cambrian nomenclature can be considered well established; additional study is necessary before details of the stratigraphy are available. The Cambrian in the Big Horn Mountains appears to be of the same general character as the sequence in the Bridger Mountains, is roughly equivalent to the Depass and Gallatin formations, and it is probable that these formational names can be applied in at least part of the range. However, in most of the Cambrian areas in the Big Horns, by far the greater part of the beds is covered by mantle and glacial deposits, and it will be difficult to separate the sequence into formations.

RATTLESNAKE–CEDAR MOUNTAIN UPLIFT

The Cambrian sequence in the Shoshone Canyon, 7 miles west of Cody, is divisible into the Flathead, Gros Ventre, and Gallatin formations. The section is similar to the section at Crow Creek Canyon, in the western part of the Owl Creek Mountains; and the Teton Canyon and Du Noir members both are present. The upper part of the Gallatin contains black and gray shales carrying graptolites, some of which have been described by Ruedemann.[17] The writer has not found these graptolite shales in any other section.

AGE AND CORRELATION OF THE FORMATIONS

The correlation of the Cambrian formations of northwestern Wyoming with the Cambrian of other regions must be based on evidence that is far from complete. The faunal lists given in the preceding section of this paper represent only a small part of the fauna that ultimately will be listed from the beds, inasmuch as the collections include species unidentified because of their fragmentary condition. Many of the species and genera found were undescribed;

[17] Rudolf Ruedemann, "The Cambrian of the Upper Mississippi Valley," Part III, "Graptolitoidea," *Public Mus. of the City of Milwaukee Bull. 12*, No. 3 (1933), p. 323.

and although the writer is publishing some of them in connection with this study, only those which can be closely compared to described forms are of value in correlation. Furthermore, many of the forms which were described from the Cambrian of other regions and which also occur in Wyoming have never been accurately tabulated as to their precise stratigraphic position; thus their use in correlation is impaired. In addition, there is difficulty in that the Cambrian of the Pacific Province has not been adequately subdivided into standard faunal zones, and this is accentuated by the fact that the type regions for the Waucobian, Albertan, and St. Croixian series, representing Lower, Middle, and Upper Cambrian, respectively, are in widely separated parts of the continent. For these reasons the correlations proposed in this paper are considered to be only approximate.

The correlations are based on an evaluation of the trilobite faunas occurring in the formations, together with a consideration of the sequence of the faunas and their positions in the sections. The only other organisms represented in any abundance are the brachiopods, and these are simple in type and apparently not diagnostic. The correlations are based only on those genera and species whose range in Cambrian rocks outside of Wyoming appears to be well established; the associated forms are listed in the tables in the preceding section of this paper.

Correlation of the Gallatin formation.—The most reliable correlations which can be made between the Cambrian formations of northwestern Wyoming and the Cambrian of other regions are afforded by a comparison of the faunal zones of the Gallatin with zones in the type St. Croixian of the upper Mississippi Valley (Fig. 6).

The Du Noir member of the Gallatin contains a fauna characterized by *Kingstonia*, several species of *Crepicephalus* belonging to the *C. texanus* group, *Blountia*, and *Cedaria*. So far as known to the writer, all the described species belonging to these genera occur in beds of early St. Croixian age. In the upper Mississippi Valley, *Crepicephalus* and *Cedaria* characterize the Eau Claire member of the Dresbach; and the Du Noir is thus believed to be approximately of Eau Claire age. In the Black Hills, *Crepicephalus* and *Kingstonia*

occur in a zone near the base of the type Deadwood;[18] for this reason, the base of the Gallatin and the base of the Deadwood appear to be approximately of the same age and the Wyoming "Deadwood" evidently includes beds considerably older than the type Deadwood.

The succeeding faunal zone of the Gallatin is characterized by *Elvinia, Camaraspis,* and *Pterocephalia.* In the type St. Croixian, *Elvinia* and *Camaraspis* occur in the Ironton member of the Franconia. The beds immediately overlying the *Elvinia* zone of the Gallatin contain *Irvingella gibba* and *I. major;* the latter species characterizes the transition beds between the Ironton and upper Franconia.[19] Therefore, it is thought that the part of the Gallatin between the Du Noir member and the *Irvingella* zone corresponds approximately to the Galesville (upper Dresbach) and the Ironton.

The *Taenicephalus* zone of the Gallatin (Fig. 6) contains *Taenicephalus cordillerensis* and *Wilbernia* sp. aff. *W. diademata,* both of which are closely related to species occurring in the *Conaspis* beds of the Franconia. The beds immediately succeeding the *Taenicephalus* zone contain *Idahoia,* which was first described from the Ovid formation, of Oneida County, Idaho; Resser has recently referred several species occurring in the Franconia to *Idahoia.*[20]

The correlation of the beds overlying the *Idahoia* zone is uncertain. The only significant trilobite found is *Briscoia schucherti,* which occurs 22 feet below the top at Du Noir; this species was originally described from the "Mazomanie," the uppermost member of the Franconia.

Correlation of the pre-Gallatin formations.—The writer has found few well-preserved, useful fossils in the pre-Gallatin formations. None of these seems to occur in persistent faunal zones, a number of them were undescribed, and it is impossible at present to correlate accurately the pre-Gallatin formations with the Cambrian formations of other regions.

No diagnostic species have been found in the Flathead formation

[18] Meyerhoff and Lochman, *loc. cit.*

[19] J. M. Wanemacher, W. H. Twenhofel, and G. O. Raasch, "The Paleozoic Strata of the Baraboo Area, Wisconsin," *Amer. Jour. Sci.,* 5th ser., Vol. XXVIII (1934), p. 6.

[20] C. E. Resser, "Nomenclature of Some Cambrian Trilobites," *Smith. Inst. Misc. Coll.,* Vol. XCIII, No. 5 (1935), pp. 35–36.

or in the lower shales of the Gros Ventre below the Death Canyon member. The total fauna identified by the writer from the Death Canyon is composed of *Bolaspis* (?) *resseri*, *Marjumia* (?) *tetonensis*, *Glyphaspis* sp. cf. *G. perconcava*, *G.* sp. und., and aff. *Kochaspis* sp. und. All of these species, except the questionable *Marjumia*, are sufficiently similar to known forms to admit of no doubt of their Middle Cambrian (Albertan) age; *Marjumia* (?) *tetonensis* possibly represents a previously unknown genus. All of the species occur in the upper part of the Death Canyon; and all were found in the Teton and Gros Ventre Mountains except the questionable *Kochaspis*, which was found in the western part of the Owl Creek Range.

A small trilobite fauna was found in the upper shales of the Gros Ventre, about 120 feet above the Death Canyon member, in the western Owl Creek Mountains. The fauna includes *Marjumia* (?) *tipperaryensis*, representing the same genus as the questionable *Marjumia* from the Death Canyon member in the Teton Mountains, and *Crepicephalus* (?) *wyomingensis;* the latter species is referred to *Crepicephalus* because of its similarity to *C. coosensis* Walcott, which occurs in the Middle Cambrian of Alabama and probably is not a *Crepicephalus*. The writer has found no other diagnostic fossils in the upper Gros Ventre shales, although in the collections at the United States National Museum there are Middle Cambrian trilobites which came from the upper shales of the Gros Ventre, between the Death Canyon member and the Gallatin, in the Teton Mountains.[21]

The paleontologic evidence indicates that the Gros Ventre and Flathead are largely, if not wholly, of Middle Cambrian age; and although no diagnostic fossils have been found in the Depass, its stratigraphic relations with the Gallatin and Gros Ventre indicate that it is largely Middle Cambrian (Fig. 5). It seems also that the pre-Gallatin beds must include late Middle Cambrian, inasmuch as through most of the Wind River Mountains and the Owl Creek–Bridger uplift they appear to be conformable below the Gallatin, which has an early Upper Cambrian fauna at its base. However, as previously pointed out (p. 122), there are about 75–100 feet of beds, mainly limestones, which are present in the upper part of the Gros

[21] C. E. Resser, personal communication.

Ventre in the northwestern Wind River Mountains and in sections to the east, but which are largely absent in the Gros Ventre and Teton ranges, where they seem to be represented by a disconformity. Furthermore, the Du Noir member of the Gallatin contains an Eau Claire fauna; and therefore it is assumed that equivalents of the Mount Simon, which underlies the Eau Claire and contains no fauna, are represented in the Wind River, Owl Creek, and Bridger ranges by approximately the upper 75 or 100 feet of the pre-Gallatin beds (Fig. 6).

CAMBRIAN HISTORY AND PALEOGEOGRAPHY OF WYOMING

During early Cambrian time most, if not all, of Wyoming was undergoing peneplanation. No one, to the writer's knowledge, has demonstrated the presence of Lower Cambrian rocks in Wyoming, although it is known that the early Cambrian sea was present in Utah, and it is considered possible by the writer that this sea extended into southwestern Wyoming, as indicated by Schuchert on a recent paleogeographic map.[22]

In northwestern Wyoming the record begins with the advance of a sea which reached the western border of the state in later Middle Cambrian time and spread slowly eastward. In this sea were deposited the Flathead sandstones and the greater part of the shales and limestones which compose the Gros Ventre and Depass formations, the beds overlapping toward the east (Fig. 5). Toward the close of Middle Cambrian time an uplift took place along the northwestern boundary of the state, the seas retreating from the area west of the Wind River Mountains but remaining over the greater part of the area to the east, where deposition appears to have continued into the later Cambrian with little or no interruption.

During later Cambrian time the sea returned to the area that had previously emerged and spread farther toward the east, extending across Wyoming at least as far as the Black Hills and probably connecting eastward with the upper Mississippi Valley sea. The sea remained in Wyoming through the greater part of later Cambrian time, although there were irregular withdrawals, as indicated by disconformities in the upper part of the Gallatin formation; and it is

[22] Charles Schuchert and Carl O. Dunbar, *Textbook of Geology*, Part II, *Historical Geology* (New York: John Wiley & Sons, 1933), p. 125.

probable that the sea retreated completely before the close of the period.

The history of central Wyoming during the Cambrian period is uncertain. The Cambrian rocks are known to thin and become increasingly clastic from the Wind River Mountains southeast to central Wyoming, where the beds have been referred by various writers to the "Deadwood" formation; they disappear in the northern part of the Laramie Range between the Casper and Douglas districts.[23] In the Rattlesnake Mountains in the eastern part of Natrona County, the Cambrian consists of 800 feet of sandstone, shale, and impure limestone; at Alcova, 25 miles southeast of Casper, there are 200 feet of sandstone, quartzite, and conglomerate; and at Rawlins, 65 miles south of the Rattlesnake Mountains, 395 feet of quartzite and conglomerate.[24] At all of these localities the Cambrian is overlain by the Madison limestone (Mississippian), the Big Horn formation terminating at the southeastern end of the Wind River Range.[25]

The character of the Cambrian of central Wyoming suggests nearshore deposition, from which it is inferred that the Cambrian sea did not extend to southeastern Wyoming, where Cambrian rocks are absent. It is the writer's opinion that the thinning of the beds is due in large part to overlap, although it is probable that the thickness has been somewhat reduced by erosion prior to the deposition of the Madison. If this interpretation is correct, southeastern Wyoming was a part of the land mass, Siouxia, which acted as the dominant source of the Cambrian sediments which were deposited to the northwest.

ACKNOWLEDGMENTS.—The writer wishes to express his appreciation to Dr. G. M. Kay for his many suggestions and helpful criticisms. The suggestions of Dr. H. N. Coryell also have been appreciated. Mr. A. G. Lucier aided in the photographic work, and Mr. James Kindler acted as camp assistant during part of the summer of 1934.

The field studies were carried out as part of the program of the Wyoming Geological Survey, and the writer is indebted to Dr. S. H. Knight for making available the resources of that organization.

[23] Willis T. Lee, "Correlation of Geologic Formations between East-Central Colorado, Central Wyoming, and Southern Montana," *U.S. Geol. Surv. Prof. Paper 149* (1927), p. 46.

[24] *Ibid.*, pp. 45, 51, 73.

[25] N. H. Darton, "Paleozoic and Mesozoic of Central Wyoming," *Geol. Soc. Amer. Bull. 19* (1908), p. 410.

SUCCESSIVE PEDIMENTS AND TERRACES OF THE UPPER RIO PUERCO IN NEW MEXICO

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ABSTRACT

In the area drained by the Upper Rio Puerco and its tributaries, erosion has been marked by the development of a series of successively lowered, stabilized, local base-levels whose existence is recorded by multiple pediments and terraces. The early geomorphologic history is obscure. Some indications of a hypothetical early and widespread erosion surface exist. There are more definite traces of a later high-level surface of coalescing pediments. The detailed history began in the Pleistocene with the dissection of this second surface and involved formation of the Upper Rio Puerco by adjustment to structure with accompanying piracy. Thereafter, the La Jara pediment was formed to a stabilized grade 180–200 feet above that of the present stream. Renewed incision and adjustment to structure occurred and the Rito Leche pediment was formed to a stabilized grade 75–85 feet above the present stream. Two minor cycles produced terraces about 30 and 10 feet high, respectively, but on rock. Finally, the inner valley of the river was cut and alternately filled and eroded in the minor epicycles characteristic of the region.

LOCATION OF THE AREA

The area here considered is the valley of the Rio Puerco at the base of the Nacimiento Uplift in central New Mexico extending about 10 miles north and 30 miles south (Fig. 1) of the village of Cuba, New Mexico. It marks the boundary between the Plateau Province and the Western Prong of the Southern Rocky Mountains. This portion of the Rio Puerco north of the village of Cabezon or, more precisely, the mouth of the Arroyo Chico, is here called "Upper Rio Puerco," and has a somewhat different history from the "Lower Rio Puerco" which forms in part the boundary between the Plateau and the Basin and Range provinces.

For an intermittent and in part ephemeral stream the Rio Puerco is unusually long, and with the exception of the Pecos is the largest tributary of the Rio Grande in New Mexico. From Cuba it flows nearly due south 140 miles to its junction with the Rio Grande at La Joya. Just north of Cuba it is formed by the union of two streams each about 10 miles long. The Arroyo San José, which flows from the

north parallel to the uplift and along the axis of the main stream, should be considered its true extension upstream. At Cuba the Arroyo San José is joined by a stream that flows from the east out of the Nacimiento Uplift, locally called San Pedro Mountain, through the Vallecito del Rio Puerco. Because this stream carries the largest low-water flow, it bears the name "Rio Puerco." The Arroyo Chico is an important tributary of the Rio Puerco, entering it from the west 103 miles from its mouth and 2 miles south of the town of Cabezon. Here is the dividing mark between the upper and lower parts of the Rio Puerco.

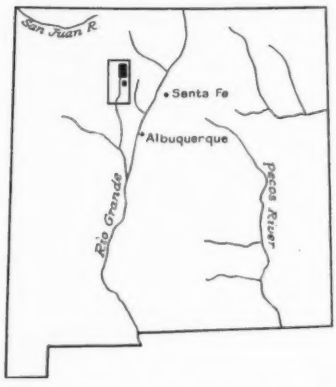

FIG. 1.—Index map showing location of Figure 2 in outline and Figures 4 and 7 in black.

RELATION OF STRUCTURE AND TOPOGRAPHY TO PEDIMENTS

The Upper Rio Puerco, including the Arroyo San José, flows approximately parallel to and just west of the long, north-south axis of the Nacimiento Uplift. Its valley is bordered on the west by sandstone cliffs and on the east by gravel-capped ridges and spurs, the remnants of once extensive pediments cut by its eastern tributaries. It flows in greater part in broad valleys on a well-developed floodplain in belts of soft shale where conditions are suitable for the development of nearly perfect pediments, but in some parts it occupies narrow, canyon-like valleys cut in sandstone, a situation which favors the formation of terraces rather than pediments. Such a canyon lies just north of La Ventana where the stream cuts the resistant Mesa Verde sandstone. Similarly, above Cuba the Arroyo San José flows through a narrow canyon where it crosses the sandstones of the upper Nacimiento beds and basal Wasatch formation. In these localities there is little trace of pediments, which are normally cut by side streams, but the gravel-capped terraces of the main stream are well preserved.

GENERAL NATURE OF THE PROBLEM

It is not the purpose of this paper to review the extensive literature on the typical erosion surfaces of arid regions, now generally known as pediments. Here it is proposed only to describe the erosion surfaces of a particular area in central New Mexico, to outline the physiographic history, and to present an analysis of the characteristic topography. Only incidentally are there references to the processes involved or to the general problems of central New Mexico, which will be considered more fully in future papers by the writers and their colleagues.

The study of pediments in this area presents an intricate problem. The pediments have all been thoroughly dissected, and the lower portions of the higher surfaces, in places, have been almost completely destroyed. Thus the determination of the number of erosion surfaces represented by the existing remnants is rendered difficult. In some localities the evidence is so meager or conflicting that the authors present their conclusions as merely tentative. However, within the area of the Gallina topographic sheet of the United States Geological Survey mapping of the existing remnants of each pediment afforded opportunity for a continuous check on the number and relative heights of the several ancient surfaces. The topographic detail shown on this map is somewhat inadequate because of its small scale (1:125,000) and its large contour interval (100 ft.). Mapping on a larger scale would have lent greater facility and increased validity to the work.

Mapping required that distinction be made in the field between main-stream terraces and side-stream pediments, and made necessary the determination of the relative position of the Upper Rio Puerco at the time of cutting of any particular pediment. The complications involved are discussed in a later paragraph. Also a description of the kind of rock beveled, the nature of side slopes of pediment remnants, and thickness and composition of the gravel cover will be given for typical areas.

A truly adequate map of the pediments of an area should show both the undissected remnants of each original surface and the slopes leading down to the next lower surface. This would require a large-

scale base map and two patterns or colors for each pediment. In the areas mapped, not only was the scale of the base maps inadequate but their generalized nature precluded accurate location in the field. Also economy in printing required maps in black and white. Hence in Figures 4 and 7 each pattern represents the undissected remnants of a pediment and the slopes leading down to the next lower surface. The pattern for residuals applies to residuals above the highest surface except for a few isolated hills that may be merely residual above the next lower surface.

OUTLINE OF DEVELOPMENT OF PEDIMENTS

Under continually rising local baselevel.—In many of the discussions of the development of pediments, a continually rising local baselevel is assumed as the only possible basic condition.[1] As this condition is restricted to those enclosed basins of arid and semi-arid regions which have no through drainage, and, even within these basins, is confined to areas near the site of deposition, it is comparatively rare. An enclosed basin can be formed only if uplift occurs at a rate sufficient to break up and disintegrate the original drainage. It can endure only so long as reintegration is prevented either by continued uplift or by the resistance of the uplifted masses to their reduction by erosion. All known examples of such basins have originated through normal faulting. As soon as the basin is formed, it begins to be filled with detritus derived from the surrounding higher blocks. As these higher blocks are consumed by erosion, their steep walls retreat, with the formation of a sloping plain at their base. Regardless of the process by which this plain is formed, its local baselevel is the edge of the alluvial fill of the basin. This feather edge, rising all the while, advances toward the mountain blocks, and, consequently, by successive increments, a convex-cut surface, which Lawson has called the "sub-alluvial bench," is produced and almost immediately buried by the ever encroaching alluvium. The processes of planation vary in relative importance from place to place, one being dominant here, another there. Throughout the cycle of development the rate of rise of the alluvial fill diminishes, because,

[1] As a recent example, see J. H. Maxon and A. H. Anderson, "Terminology of Surface Forms of the Erosion Cycle," *Jour. Geol.*, Vol. XLIII (1935), pp. 88–96.

at the beginning, the basin has a smaller volume for each unit of depth and because, toward the end, the surrounding highlands are so reduced as to furnish detritus in diminishing quantities. Eventually the rate of rise becomes so slight and the advance of the feather edge of the alluvium is so slow that extensive pediments develop. Much of the recent discussion of pediments has to do with the relative strength of the several processes of planation under the foregoing conditions. Generally speaking, lateral planation is most important until the late stages of the cycle when, because of the decrease in water and material, the others take precedence.[2]

The pediment has a surface which is concave upward, increasing in steepness of slope toward the mountains. It is covered with a mantle of débris in transit, which is thin at the margin and thickens toward the center of the basin. The rising local baselevel tends to induce a somewhat thicker cover on the pediment than would accumulate with a perfectly stabilized baselevel.

Under stable local baselevel.—The possibility of a stable local baselevel appears, at first glance, to be a remote or purely theoretical contingency. In many arid and semi-arid basins, however, the local baselevel is a through-flowing stream which has a graded course. At any one point in its course and over long periods of time it may fluctuate so little in elevation that it is, for all practical purposes, a stable local baselevel. The adjoining areas will be eroded by ephemeral streams to sloping plains or coalescing pediments graded to this baselevel. The fronts of nearby mountains will probably retreat in much the same fashion as if they bordered a basin possessed of a constantly and rapidly rising baselevel. Superficially the sloping plains between mountain blocks and the river will resemble the plains of enclosed basins. However, the convex alluvium-covered surface, "sub-alluvial bench" of Lawson, will be absent. The thickness of the alluvial cover will be moderate. Beneath the floodplain of the graded river and beneath the channels of the correspondingly graded tributaries, it will extend only to the depth of possible scour and fill.

In the Basin and Range Province the establishment of such a local

[2] For fuller discussion see Kirk Bryan, "The Formation of Pediments," *Rept. XVI Intl. Geol. Cong.*, preprint, 11 pp.; November, 1935.

baselevel may occur after the original enclosed basins have been integrated and have become parts of an area draining to the sea as outlined deductively by Davis.[3] In such an area that part of the surface corresponding to the original filled basins is underlain by great thicknesses of alluvium, whereas adjacent parts are cut on the rocks of the mountain masses. The cut surface consists of coalescing pediments that may consist of bare rock or of rock only thinly mantled by detritus in transit. Outward from the mountains, however, these pediments merge abruptly with alluvial plains underlain by great depths of alluvium. If the rate of fill of the basins is rapid, the thick alluvium rests on a structural floor as shown by Bryan in 1922.[4] If the rate of fill is more gradual, especially toward the end of the process, the alluvium will rest partly on a structural floor and partly on a "sub-alluvial bench."[5]

The Basin and Range Province in central New Mexico has a more complicated history than is usually postulated in the hypothesis of continually rising baselevel or in Davis' argument in regard to basin integration. In Miocene time basins and mountain blocks were formed apparently by block faulting. The fact of the formation of the basins and ranges is well established, but whether block faulting was the sole process remains to be established. These basins were filled by the deposition of the Santa Fe formation of early Pliocene or late Miocene age. Renewed faulting uplifted the old and also a few new mountain blocks and initiated the Pleistocene period. Thereafter integration of drainage occurred and the Rio Grande became the master-stream with a stabilized grade which, near Albuquerque, was about 500 feet above its present grade. Extensive erosion surfaces were cut by all the tributaries. These surfaces formed great coalescing pediments on the hard rocks of the mountain blocks and on the deformed and only slightly consolidated late Tertiary Santa Fe beds.[6]

[3] William Morris Davis, "The Geographical Cycle in an Arid Climate, *Jour. Geol.*, Vol. XIII (1905), pp. 381–407.

[4] Kirk Bryan, "Erosion and Sedimentation in the Papago Country, Arizona," *U.S. Geol. Surv. Bull. 730* (1923), p. 59 and Fig. 16. This work was first published as a separate in 1922.

[5] *Ibid.*, pp. 61–62, Fig. 20.

[6] This sequence of events is considered more in detail in a paper on the Ceja del Rio Puerco, now in preparation.

The area here considered is, however, part of the Plateau Province and is unaffected by faulting of later Tertiary time. So far as now determined its latest deformation is the post-Eocene uplift of the Nacimiento Range. With respect to the adjacent Basin and Range Province, this area is a bordering highland and its history is similar to that of one of the mountain blocks within the Basin and Range Province. Erosion has proceeded with respect to local and temporary baselevels established outside of the area. Throughout the Pleistocene period these exterior but controlling baselevels were alternately stabilized for long intervals of time and then lowered rapidly. These alternations occurred with amplitude decreasing both in the length of the period of stabilization and in the amount of lowering.

Under successively lowered stable local baselevels.—After the formation of extensive pediments graded to a through-going stream whose gradient has been for a long period nearly stable, the cycle may be interrupted. Both main stream and side streams entrench themselves below the coalescing pediments just developed. If, thereafter, the main stream becomes graded, it again forms a local, stable baselevel. Unless the second pediment be less extensive than the first, all trace of the earlier one will be obliterated. If, as the tributaries attain grade, the second surface is formed as a pediment that is less extensive and has a lower elevation, the remnants of the first pediment will remain as buttes and flat-topped ridges between adjacent side streams. At the base of mountains the ridges may be attached to the higher mass between canyon mouths and will extend outward as sloping spurs. But subsequent streams may convert them into isolated buttes and ridges surrounded by the lower and continually enlarging erosion surface. Generally these ridges will be capped by the gravel which in some localities is the characteristic cover of pediments. If the cover is moderately thick, it may serve as an effective protection against the destruction of the remnant by erosion.

The incision of the through-flowing river and its renewed stabilization may be brought about by either regional or local uplift or by climatic change. The Upper Rio Puerco discharges into the Lower Rio Puerco and then flows more than 100 miles to the Rio Grande. The Rio Grande forms the local baselevel, and it has been subject

throughout Pleistocene time to alternate periods of incision and of stabilization of grade. The ultimate cause of this characteristic of the Rio Grande cannot be determined until the relation of river to sea-level throughout Pleistocene time is known and the location and amount of Pleistocene uplift in Texas and New Mexico is better understood.

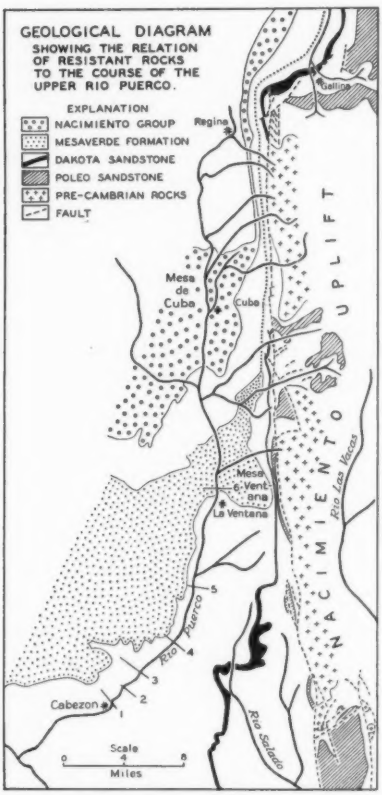

FIG. 2.—Geologic diagram showing the relation of resistant rocks to the course of the Upper Rio Puerco.

SKETCH OF THE GEOLOGY

The distribution and extent of the terraces and pediments on the western flank of the Nacimiento Uplift are controlled by the structure and stratigraphy of this range. The following résumé of the geology of this uplift is based largely upon the excellent work of Renick.[7] His results have been checked and, in certain details, corrected by the Harvard Field schools of 1931 and 1932.

The Nacimiento Uplift is an asymmetric, domal anticline, about 50 miles long from north to south, bordered on the west by a north-south high-angle thrust fault. There are places where this fault becomes a zone of faults, and at both ends of the uplift faulting is much complicated (Fig. 2). The maximum stratigraphic displacement is 3,000–3,500 feet. The overturned sedimentary beds on the down-

[7] B. C. Renick, "Geology and Ground-water Resources of Western Sandoval County, New Mexico," *U.S. Geol. Surv. Water Supply Paper 620* (1931).

thrown, western side flatten out so rapidly that a hundred yards west of the fault the dip may be as low as 5 degrees.

The core of the uplift is a pre-Cambrian basal complex, mostly gneissic granite, with minor areas of diorite and schist. Although only a small area of quartzite exists at the north end of the range, it seems probable that this rock was formerly more extensive, for all of the Mesozoic conglomerates contain quartzite pebbles. The basal complex is, under present conditions of erosion, the most resistant rock unit in the area. The Paleozoic rocks include the Magdalena, mostly limestone, and the Abo and Chupadera, which are sandstone with minor quantities of shale. Both the Abo and Chupadera are red or reddish and contain conglomeratic zones which have furnished pebbles of quartz, quartzite, and chert to later formations. For the most part these rocks crop out east of the fault and, because of their resistance to erosion, form part of the mountain. In only a few places are they sufficiently reduced to be included in the area of pediments.

The Mesozoic begins with the Poleo, a cliff-forming sandstone which contains at its base a conglomeratic zone similar to that of the Abo and Chupadera. Above it is the much less resistant red Chinle shale and sandstone which usually forms valleys. The Wingate sandstone, capped by Todilto gypsum, forms minor cliffs. Above lie the easily eroded, variegated shales and sandstones of the Morrison formation.

The Cretaceous consists of massive, buff sandstone and great thicknesses of dark shale. Its topography is characteristic and dominates the area between the Nacimiento Uplift and the Rio Puerco as well as a large part of the San Juan country to the west. The sandstones, which range in thickness from 100 to 1,000 feet and in general are 200–400 feet thick, form escarpments, low buttes, and mesas. The shales, which range from 1,000 to 2,000 feet in thickness, form extensive plains and lowlands. In this eastern part of the San Juan structural basin the usual dip of the rock is about 2 degrees, and the broad lowlands between the sandstone escarpments are 10–15 miles across. Even close to the Nacimiento Uplift, where the dips may range from 45 to 90 degrees, the thick shales form valleys three to four times as broad as the narrow sandstone ridges and escarpments.

Altogether there are three alternations or couples of sandstone and shale. The Cretaceous, as strictly defined, consists of two sandstone formations, the Dakota and Mesa Verde, and two shale formations, the Mancos and the Lewis. The doubtfully Cretaceous formations, the Ojo Alamo sandstone and the Kirtland-Fruitland group, together with the Paleocene Nacimiento (Puerco-Torrejon) group, form another sandstone and shale combination which continues the type of topography characteristic of the true Cretaceous.

The Eocene is represented by the Wasatch, which is characterized by a basal conglomeratic sandstone and great thicknesses of variegated, sandy shale which, where not planed down into pediments, rise out of the plains as isolated and picturesque buttes.

The Nacimiento Uplift lies on the southeastern margin of the San Juan structural basin, with indications that the basin was formed first. The lowest part of the structural basin lies to the northeast of its geographical center and only a short distance northwest of the northern end of the Nacimiento Uplift. Throughout the area here considered the rocks of the basin retain a northeast strike and a low northwest dip (2°–5°) until they are bent up suddenly along the north-south trend of the uplift. This relationship is shown in Figure 3, and it is evident that it is the northeast strike characteristic of the rocks of the San Juan Basin which has largely controlled the subsequent course of the Rio Puerco and its tributaries.

ORIGIN OF THE UPPER RIO PUERCO

Early surfaces.—The series of pediments and rock terraces of the Upper Rio Puerco have developed under successively lowered, stable, local baselevels. However, the main stream which formed the local baselevel has shifted in position on account of piracy, and consequently there have been changes in local baselevel not entirely consistent with simple, successive lowering.

Prior to the existence of the Upper Rio Puerco the area was drained by another stream, here called the Rio Chacra, which flowed southeast instead of south, crossing the present course of the Rio Puerco near the Mesa Prieta. Several successive piracies transformed this ancient stream into the Rio Puerco. The Rio Chacra and its tributaries probably flowed on an extensive erosion surface of

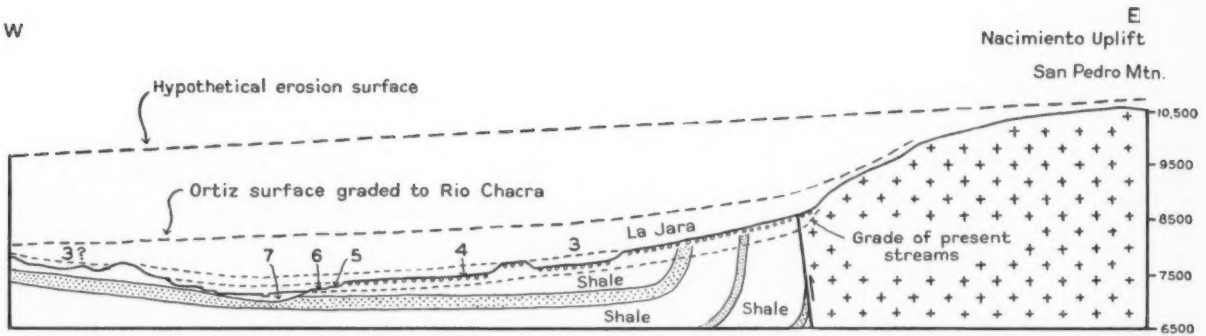

FIG. 3.—Ideal geologic section showing relation of structure to erosion surfaces

coalescing pediments, which extended from the San Juan region southeast between the Nacimiento Uplift and Mount Taylor. The streams draining the highlands of the Nacimiento Uplift had a westerly course toward the Rio Chacra and flowed over an extensive sloping surface. There was a similar slope north and northeastward from a highland at Mount Taylor, where this extensive erosion surface, cut on the Cretaceous rocks, is overlain by the lava of the present Sierra Chivato or Mesa Chivato. Similarly, lava flowed down the central depression of the Rio Chacra and this part of the surface still survives beneath the lava cap of the Mesa Prieta. The course of the ancient Rio Chacra and its relation to this extensive erosion surface (Ortiz surface) are further considered in a forthcoming paper on the Ceja del Rio Puerco, one of the most extensive remnants of this surface.[8]

In the Nacimiento Uplift, evidence of the existence of two high-level surfaces is preserved. The top of San Pedro Mountain at the north end of the uplift is a gently rolling surface of more than 30 square miles (Fig. 4). It has elevations of 10,000–10,500 feet or more. From this essentially flat surface long, sloping spurs descend gently to elevations of 9,500–9,000 feet on the west, and to 9,000–8,500 feet at the north end of the uplift. Below this rounded slope the spurs break off sharply to the lower pediments described below. A few miles northeast of Cuba, Renick[9] reports gravels mantling the spurs at elevations as high as 8,600 feet; according to the observations of the writers these gravels are obviously older than the pediments described in this paper which intersect the mountain front at elevations 200–300 feet lower. The long spurs appear to represent all that is left of the once extensive Ortiz surface.

The upper surface of the San Pedro Mountain (Fig. 4) appears to be part of a once very extensive and doubtless quite perfectly developed erosion surface whose correlation with remnants of the same age, general character, and similar antiquity in other mountains of the region is beyond the scope of this paper. There is also a very flat

[8] The term "Ortiz surface" is derived from the Ortiz Mountain of Santa Fe County whose beautifully developed pediments have been much discussed. The correlation of the surface on which the Rio Chacra flowed with these distant surfaces involves a great extent of country and is deferred to later papers.

[9] *Op. cit.*, p. 58.

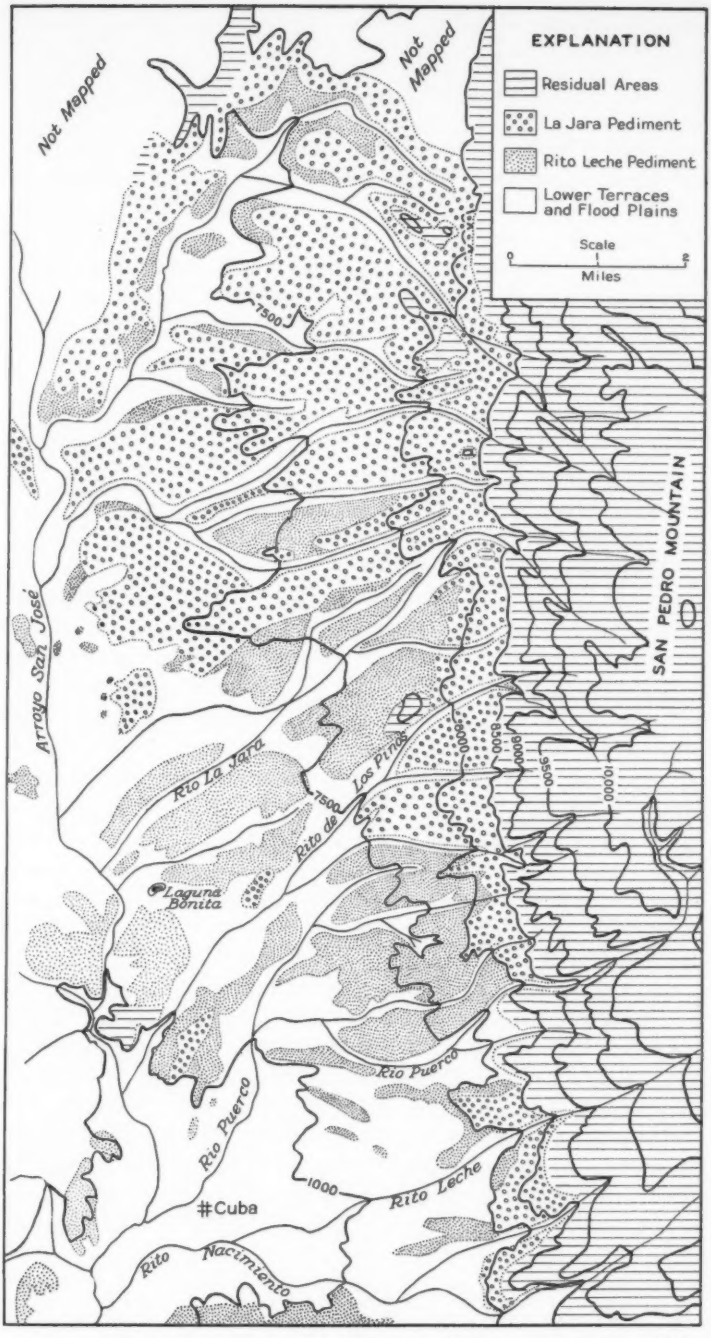

FIG. 4.—Map showing remnants of the La Jara and Rito Leche pediments near Cuba, New Mexico.

and perfect erosion surface on the Cuba–Chaco Canyon road, along the divide between the Rio Puerco and San Juan drainages at an elevation of about 7,200 feet. It bevels the underlying beds which dip gently northward and is mantled by a thick soil, partly wind-blown. Westward, spurs slope to the Rio Chaco and Rio San Juan, and southward they slope toward the supposed position of the ancient Rio Chacra. This surface, which may be called Haynes Flat, may possibly be a correlative of the surface represented by the top of San Pedro Mountain. It is here assumed that Haynes Flat was the northern divide for the Rio Chacra drainage and that the spurs sloping southward from it represent the extensive pediment (Ortiz surface) drained by the Rio Chacra. Obviously, our present knowledge is too uncertain and the maps of the region too generalized to make critical studies of these questions profitable.

Distribution of resistant rocks in the area.—The stream captures by which the Upper Rio Chacra became the Upper Rio Puerco were controlled by the distribution of the resistant rocks in the area. The Permian, Triassic, and Jurassic strata, which form minor anticlines at the southwestern base of the Nacimiento Uplift, have had little influence on the drainage development of the Rio Puerco, but have influenced the course of the Rio Salado, a tributary of the Rio Jemez. The main rock units of the Cretaceous and Tertiary are shown in Figure 2, and their course across country in relation to the drainage can be traced. Generally speaking, their strike is northeast-southwest. The Dakota, although having in general a north-south strike, has, because of its resistance, played a considerable part in drainage changes. In general the headwater tributaries of the Rio Salado are developed in the anticlines to the east of its line of outcrop. North of the minor anticlines from Los Bancos to Nacimiento Creek, it forms a hogback parallel to the main mountain mass. Each stream as it flows from the mountains emerges through a narrow gap or gateway in the Dakota ridge. Thus in miniature the mountain border resembles the east front of the Rocky Mountains in Colorado.

The resistant beds, most influential on the drainage pattern, are the sandstones of the Mesa Verde formation and those at the base of the Nacimiento and Wasatch formations. These sandstones, as shown on Figure 2, have strong northeast trends as they approach

the Nacimiento Uplift. At the time of the Ortiz surface the relief of their outcrop was so reduced that streams flowed out of the uplift westward to the Rio Chacra. Only the Mesa Ventana, carved from the Mesa Verde, and the Mesa de Cuba, carved from the Ojo Alamo sandstone, stood as residuals. Thus, west of the pre-Cambrian outcrop the country during the Ortiz stage was reduced to the condition of an almost perfectly graded surface, the dry-climate equivalent of peneplanation. With the lowering of baselevel and the beginning of dissection, adjustment to structure took place. One of the tributaries of the Rio Chacra flowed southwestward across the belt of weak Mancos shale toward the volcanic neck, Cabezon, and this favorable course gave it obvious advantages over similar tributaries which crossed more resistant formations. It lengthened headward until it captured at La Ventana the streams flowing in the Lewis shale north of the Mesa Verde outcrop and thus successively gained all the streams flowing from the highland of the uplift. The Rio Chacra ceased to exist and the Upper Rio Puerco came into being. In the same period the Rio Salado, a more direct tributary to the Rio Grande by way of Jemez Creek, also came into existence by piracy. Its headwaters are now separated from those of eastern tributaries of the Rio Puerco by a divide well below the level of the Ortiz surface.

Piracy has not ceased, although the course of the Upper Rio Puerco has changed but little, being held at Cabezon and La Ventana by the gaps cut in relatively resistant rocks. At each lowering of baselevel and each incision of streams the drainage has become less indifferent to structure and more subsequent in pattern. The tributaries show this adjustment at many localities. Considering the northeast strike of the rocks the tendency has been to shift the Rio Puerco eastward. However, the strong floods of the mountain streams dump their burden of durable granitic gravel into the floodplain of the Rio Puerco and tend to force the course of the main stream westward. Periods of lowering of local baselevel, i.e., of dissection, have occurred four times since the formation of the Upper Rio Puerco. During each of these periods of dissection the eastward tendency has predominated; during each of the succeeding periods of stability of baselevel westward tendency has been important.

Now these changes in position of the main stream affect (1) the gradients of tributaries, and (2) the direction of tributaries. Thereby the direction and degree of slope of the surfaces cut by these tributaries are affected also.

PEDIMENTS

Below the ancient high hypothetical surface and the extensive Ortiz surface already mentioned, two pediments and two terraces have been formed, each graded to successively lower levels of Rio Puerco. With the establishment of the Upper Rio Puerco during the dissection of the Ortiz surface, this river has constituted the local baselevel which has governed the formation of these later surfaces. The existing remnants of these surfaces comprise gravel-capped buttes and long, sloping, flat-topped ridges that are striking features of the landscape (Figs. 5 and 6). The evidence of the existence of these ancient surfaces may be most evident where the original surface has been largely destroyed and only a ridge or a butte projects above the badlands and floodplains of the area. The gravel capping of these remnants is conspicuous because the matrix of the gravels is generally a bright, brick-red in contrast to the buff or somber colors of the underlying sedimentaries.

The higher pediment, here named La Jara from the creek of that name where it is well displayed, was graded to the Rio Puerco when it stood some 200 feet higher than at present. The next lower pediment, here named the Rito Leche from the creek along which there are conspicuous remnants, was graded to the Rio Puerco when it stood some 90 feet above its present grade. The distribution of existing remnants of these surfaces within the area of the Gallina topographic sheet, is shown in Figure 4.

The terraces are rather vaguely developed on the tributary streams, but on the Arroyo San José and on the Rio Puerco there are sharply defined benches 10–12 and 25–30 feet above the floodplain. Each terrace is a cut surface capped by 5–20 feet of gravel.

La Jara pediment.—The La Jara pediment was once an extensive surface. The tributaries of the Rio Puerco from the east, rising in the well-watered heights of the Nacimiento Uplift, had a relatively strong flow and carried a characteristic granitic gravel. These streams obliterated nearly all projections above their grade west of

FIG. 5.—View northeast toward Vallecito del Rio Puerco, showing remnants of the Rito Leche and La Jara pediments. Note that flat top of prominent ridge in middle ground (Rito Leche pediment) cuts structure of Nacimiento beds.

FIG. 6.—View north along the mountain front near San Miguel Canyon showing monadnock of Mesa Verde sandstone rising above La Jara pediment.

the mountain front. The Mesa de Cuba, capped by the basal sandstone of the Wasatch, and the Mesa Ventana, capped by massive sandstones of the Mesa Verde, were the only considerable residuals. Here and there small sandstone pinnacles projected above the graded plain as shown in Figure 6, where a part of the nearly vertical Mesa Verde forms a residual. South from San Pablo Canyon for 10 miles the nearly vertical Dakota sandstone formed a series of residuals (Fig. 7).

The western tributaries of the Rio Puerco are less numerous than the eastern tributaries. They rise in sandstone and shale plateaus of lower elevation and rainfall. The porous character of the soil leads also to lower rates of runoff. Thus they are less able to produce smooth, graded plains. However, there are at the head of the Arroyo Chihuile, the principal western tributary, broad valleys which probably represent the La Jara surface. Definite correlation has not been made, however, owing to the lack of topographic maps and to the difficulties in detecting and tracing the erosion surfaces. The streams of these Cretaceous areas carry large quantities of clay, silt, and sand, but only small quantities of gravel. The gravel consists of fragments of sandstone, iron and lime concretions, and pieces of chert and petrified wood. Slight weathering suffices to convert a stream deposit containing even fairly large gravel into a mass of clay and sand strongly resembling any other residual soil. Thus it is difficult to trace old erosion surfaces by the remnants of their overlying gravels.

The granitic gravel of the eastern tributaries is readily traced by its distinctive lithology and generally bright red color. Near the mountain front the gravel is little waterworn and contains boulders from 6 inches to 2 feet in diameter. Three or four miles from the mountains it is more rounded, 3–4 inches in diameter, and contains much sand. North of Cuba the Arroyo San José is the main stream, but it is little longer than the eastern tributaries and has a similar intermittent flow. The Rio Puerco from Cuba to La Ventana has a somewhat greater and steadier flow, but its gravels are only a little more rounded than those of the side streams. Throughout this area it is, therefore, difficult to distinguish between gravels of the tributaries and those of the main stream; hence this criterion cannot gen-

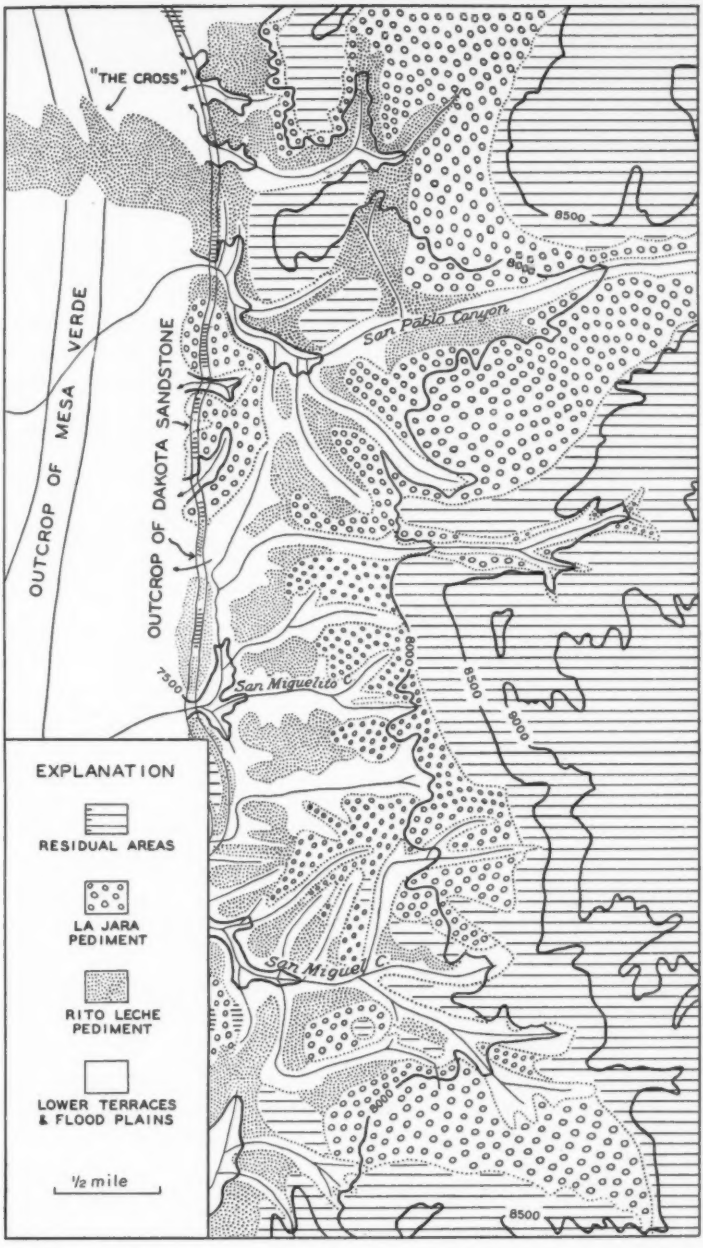

FIG. 7.—Map of high-level valleys of the La Jara and Rito Leche stages southeast of Cuba, New Mexico.

erally be used to distinguish surfaces cut by one or the other of these agents.

From La Ventana to Cabezon the Rio Puerco is a relatively large stream. The gravels of the terraces, although consisting largely of fragments of the crystalline rocks brought in by the east side tributaries, are relatively well rounded. Thus in this area the river gravel differs from the side-stream gravel and becomes a valid criterion as to whether a gravel-capped surface was cut by the main stream or by its tributaries.

Rito Leche pediment.—The Rito Leche surface near the upper end of the Arroyo San José is merely a terrace clinging to the side of the ridges which rise to the La Jara surface (Fig. 4). From La Jara Creek south the pediment was well developed and once extended from the Nacimiento Mountains to the Rio Puerco as a smooth, graded slope above which rose only small remnants of the La Jara surface. The surface sloped to a grade of the Rio Puerco about 90 feet above the present grade near Cuba but near the upper end of the Arroyo San José the gradation of the stream system was not complete and the Rito Leche surface is only about 50 feet above the floodplain.

Development of subsequent drainage and piracy.—The piracy by which the Upper Rio Puerco came into existence has already been discussed. Subsequent to this major piracy there have been a series of piracies at the headwaters of its eastern tributaries, recurring at each period of lowering of baselevel. Several individual instances were noted and many more suspected. The lack of detail on the base map used made investigation difficult. One instance will suffice. A short distance south of the Rito Leche east of Cuba lies a remnant of the Rito Leche pediment whose height above the present valley floor is 164 feet and, a mile down stream, only 145 feet. It is trenched by an abandoned valley running parallel to the long axis of the remnant, and cut 60 feet below the pediment level. This unnamed stream, flowing across the resistant Mesa Verde formation, was beheaded by a subsequent branch of Nacimiento Creek working headward on the weak Mancos shale. It is evident that piracy has occurred and that the present Rito Leche does not flow exactly parallel to the beheaded stream. If, as seems probable, this ancient

stream carved this part of the pediment of Rito Leche time, it is understandable that the direction of slope is dissimilar to that of the present Rio Leche and that, therefore, the apparent grade of the pediment is also dissimilar.

Now this relationship is not unusual and it is common for pediment remnants to have what appear to be capricious relations to existing streams. In preparing the map (Fig. 4), a number of localities were found in which the slope of the remnants of both the La Jara and Rito Leche pediments indicated that the streams by which they were formed flowed in directions considerably different from those of the present drainage. In other localities the slopes of the La Jara and Rito Leche surfaces are diverse. These anomalies can be attributed to piracy and to the increasing subsequence of drainage with each incision of streams. With the existing maps, however, the details of these drainage changes cannot be adequately determined or described.

In the region near San Pablo Canyon southeast of Cuba some details of these processes have been worked out as shown in Figure 7. The Forest Reconnaissance Map of the United States Forest Service was used in the field. On the scale of 2 inches to 1 mile and with a contour interval of 100 feet this map gives an excellent picture of the border of the mountains. However, it was made by pacing from section corners for distance and by determination of elevation with an aneroid barometer. The degree of precision of the map is governed by the method used. The map of the pediments here presented partakes of the errors of the base as well as such personal errors as are inevitable in tracing erosion surfaces through the tangled gulches of this area.

In the La Jara stage the front of the Nacimiento Uplift (Fig. 7) lacked about 400 feet of its present height. The upturned Dakota sandstone made a line of small knobs and a few similar knobs marked the outcrop of the Mesa Verde farther west. Broad valleys extended into the mountain mass. They had gradients of 300–400 feet to the mile but were relatively flat floored. They occupied nearly all the outcrop area of the early Mesozoic and the Paleozoic rocks. Only the crystalline rock, mostly granite gneiss, was persistently resistant. The change of baselevel and incision of streams in the

Rito Leche stage led to the development of streams parallel to the belts of softer rocks. The drainage of the upland to the east, which on the La Jara surface must have found its way across the Dakota by many shifting channelways, was concentrated into the San Pablo, San Miguelito, San Miguel, and San Miguel Mine canyons where "gates" were cut in the Dakota. East of the Dakota "gates," the Morrison, Wingate, Chinle, and Chupadera formations were largely reduced to flat-floored valleys from which the gypsum of the Todilto formation projected as ridges. West of the Dakota ridges only a few remnants of the La Jara surface remained as long sloping plateaus and ridges separated from the Dakota Ridge by subsequent valleys.

The later lowering of baselevel and dissection of the Rito Leche pediment produced even greater accentuation of this subsequent drainage. Eastward of the Dakota ridge and within the gaps or "gates," the etching out of the more resistant masses was accentuated. The slightly more resistant sandstones of the middle of the Morrison gave rise to secondary ridges. The Todilto ridges became more prominent and the outcrop of the Poleo sandstone farther east was accentuated into ridges. West of the Dakota the larger part of the Rito Leche pediment was swept away. The minor drainage lines adjusted themselves to courses paralleling the Mesa Verde. Even the slightly resistant sandstone and concretionary zones of the thick Mancos and Lewis shale formations became ridges between subsequent valleys.

At this time the so-called "Cross" was developed. This feature is spectacular when viewed from the high ridges as in the photograph of Figure 8 (also shown in Fig. 7). A dominant feature is a long, flat-topped ridge, a remnant of the La Jara surface. The top of this ridge passes without break across the nearly vertical beds of the Lewis, Mesa Verde, and Mancos formations. At right angles to this ridge is a subordinate ridge, a strike ridge, carved by later erosion from the Mesa Verde, which is here about 650 feet thick. The resistant basal sandstone makes the cross-arm of the "Cross." The middle zone of coal and shale is soft and forms a valley, but the upper sandy horizons also produce a series of much smaller ridges. The "Cross" is one of the most prominent evidences of the tendency to develop subsequent drainage in the dissection of widespread erosion surfaces like the La Jara pediment.

Later terraces.—The process of lowering of baselevel in steps, with intervening periods of stability, did not cease with the Rito Leche

FIG. 8.—View of mountain front northwest of San Pablo Canyon showing feature called the "Cross."

FIG. 9.—Broad valley of the Rio Puerco view toward the peak and town of Cabezon showing terraces.

pediment. Twice again the river grade was stabilized at successively lowered grades. This conclusion is based on the evidence presented by terraces adjacent to the Upper Rio Puerco, which can be traced downstream from above La Jara Creek to Cabezon. Each of these

terraces is cut on rock and is overlain by 5–15 feet of gravel. The remnants of one stand from 10 to 12 feet and those of the other from 25 to 30 feet above the level of the floodplain.

On the tributary streams similar terraces are visible, but ordinarily the existence of two is not evident, nor is it generally possible to distinguish remnants of one from those of the other.

All the remnants of these terraces on the main stream and on the tributaries are so small in area that no attempt has been made to show their location in Figure 4.

Valley floodplain.—The floodplain of the Upper Rio Puerco is broad, and over it at one time the floods spread widely. The channel of the river was shallow or practically absent. Since the late eighties, however, the channel has been incised from 20 to 50 feet.[10] At Cabezon, San Luis, and Cuba, dams have been built to raise the level of the water and permit irrigation with flood waters. In 1933 only the Cuba dam was in place.

The depth of the valley fill is not well known. In a number of places the present deep arroyo is cut into the bedrock. It seems likely that in most of these instances the present channel lies to one side of the center of the deep valley and has been superimposed on a rock ledge projecting from the side of the valley.

A test well drilled by the state engineer at La Jara in 1927 penetrated 51 feet of "light-yellow" sand and shale to a conglomerate which was considered by those in charge as a part of the Wasatch formation. At Cuba a similar test well went through 87 feet of material reported as "yellow and blue shale and sand" to a "light blue shale" which was considered by those in charge as part of the Puerco formation. As in these localities the Wasatch and Puerco are the expected bedrock it appears that the valley fill is, respectively, 50 and 90 feet deep. The Catron well, south of La Ventana, drilled as a test for oil, passed through 40 feet of brown adobe and 37 feet of brown sand before entering 418 feet of blue shale. It appears at this place that the alluvium is 77 feet deep.

The scattered and unsatisfactory data just reviewed indicate that

[10] Kirk Bryan, "Historic Evidence on Changes in the Channel of Rio Puerco, a Tributary of the Rio Grande in New Mexico," *Jour. Geol.*, Vol. XXXVI (1928), pp. 268–75.

the Rio Puerco once flowed on a gradient from 50 to 100 feet lower than that of the present. Thereafter it raised its grade by depositing the valley fill. However, at some time in the past it cut a channel in this fill similar to the present channel and again refilled this channel before the year 1885.[11] Thus alluviation and erosion appear to have alternated in recent time.

DETERMINATION OF RIVER GRADE IN THE EARLIER PHYSIOGRAPHIC STAGES

If detailed topographic maps were available the determination of the grade of Rio Puerco at each of its former gradients would be relatively simple. It would be necessary only to project the slopes of pediment remnants in the direction of the maximum slope to the position of the Rio Puerco at that stage. The position of the river would be shown by the convergence of these projected lines, particularly if remnants both east and west of the river could be used. The intersections of the projected grades of the tributaries plotted on a longitudinal profile would give the grade of the main stream for that stage. This is a standard procedure carried out with more than usual nicety by Matthes in the Yosemite Valley.[12]

In the area near Cuba it is almost impossible to carry out this process even roughly by eye because of successive shifts of the river due to piracy. An attempt was made to determine gradients in this area. The two younger terraces, 10–12 feet and 25–30 feet above the floodplain, represent strictly river-made features, and their elevations above river grade can be determined easily. However, the river-made portions of the La Jara and Rito Leche surfaces have either disappeared or cannot be distinguished from remnants of side-stream pediments, and the attempt was unsuccessful. Therefore, recourse to that portion of the river between Cabezon and the outcrop of Mesa Verde was necessary.

The broad outcrop of the Mesa Verde formation just north of LaVentana crosses the river nearly at right angles. The river lies in

[11] Data on this buried channel were gathered by Bryan in 1927; a brief account will be found in J. C. Stevens, "The Silt Problem," *Proc. Amer. Soc. Civ. Eng.*, Vol. LX, Part 1 (1934), pp. 1216–17. Further data as yet unpublished were gathered in 1933.

[12] F. E. Matthes, "Geologic History of the Yosemite Valley, California," *U.S. Geol. Surv. Prof. Paper 160* (1930), pp. 33–44, especially pl. 27.

a narrow canyon. River terraces belonging to all four of the early stages are present, as shown in cross section 6, Figure 10. Obviously the position of the river has not shifted in this locality during the successive incisions. The gravel-covered bench 185–205 feet above the floodplain occurs not only at the gap but is characteristically

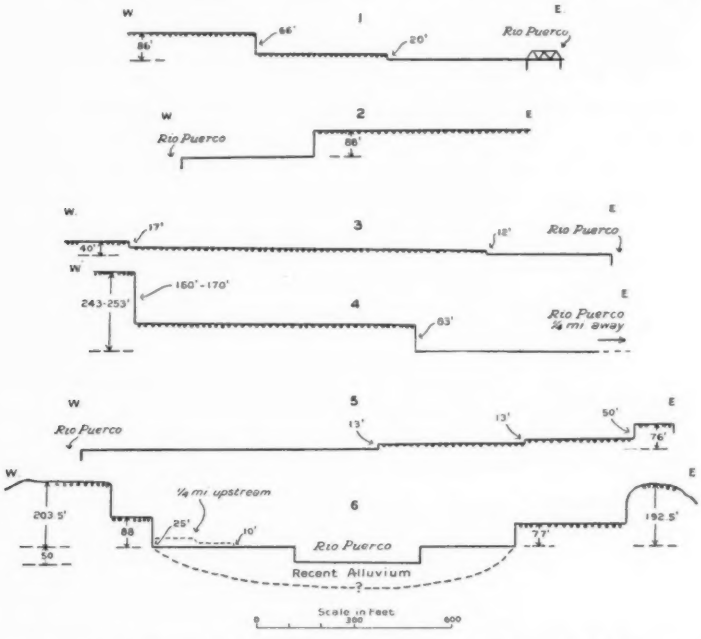

FIG. 10.—Diagrammatic cross sections of Rio Puerco Valley showing elevations of terraces.

developed on the cliffs of the Mesa Verde sandstone for about four miles in a southwesterly direction. It thus appears that since the La Jara stage the river has been shifted to the east.

The five sections of the valley southwest of this point, also shown in Figure 6, merely confirm the finding of this master section. Each of these sections was made by pacing and hand level. The fronts of the terraces were not paced and so are represented, conventionally, as being vertical, although they have a slope of 29–30 degrees. The

widths of the terraces are, however, reasonably accurate. The La Jara grade is 180–200 feet, and that of the Rito Leche is 75–85 feet above the floodplain. The two lower terraces maintain their uniform position, 25–30 feet and 10–15 feet above the floodplain.

Similar cross sections of the Cuba area were made with discordant results because of the shifts in river course and in position of side streams as previously explained. However, as already noted, the several grades become closer together near the head of the Arroyo San José. Here each successive surface is less well developed and the grades of the younger are steeper than the older. The La Jara surface is only 50–100 feet and the Rito Leche surface only 25–50 feet above the present floodplain.

SUMMARY OF PHYSIOGRAPHIC HISTORY

The nearly flat summit of the San Pedro Mountain is an ancient erosion surface whose relationships are unknown. Presumably it formed after the post-Wasatch (Eocene) folding of the Nacimiento Uplift. The surface 3,000 feet lower and 30–40 miles west, known as Haynes Flat, may be correlative or more recent. A succeeding extensive erosion surface, the Ortiz, was formed in central New Mexico and is believed to have extended into this area. This surface, on evidence from outside the Cuba area, is considered to be of late Pliocene to Pleistocene age. The long mountain spurs of the Nacimiento descending from the San Pedro Mountain and similar spurs or long ridges south of Haynes Flats may belong to it. During this stage the region was drained by a stream that has since disappeared —the Rio Chacra. With lowering of baselevel again the Upper Rio Puerco gained approximately its present course by piracy and adjustment to structure. The detailed record begins here with a less extensive surface—the La Jara pediment. During this stage the Rio Puerco had a gradient 180–200 feet higher than at present. Nearly all the area west of the crystalline rocks of the uplift was reduced to sloping plains from which rose only a few residuals.

With lowering of baselevel by 115–120 feet the river and its tributaries dissected the La Jara surface, and with stabilization of grade produced the less extensive Rito Leche surface. Twice this process was repeated to form two lower terraces, before the present inner

valley was excavated to a depth of 50 to perhaps 100 feet below the floodplain. Thereafter successive cycles of sedimentation and erosion have occurred within Recent time.

ACKNOWLEDGMENTS.—Study of land forms in the Cuba area was begun in 1931 as part of the work of the Harvard Field School in Geology, supported by the Sayles Fund. McCann spent the month of July mapping pediments and also twenty days in July, 1932. In August of the same year McCann spent an additional two weeks and Bryan three days in the field. In 1933, both authors, supported by the Milton Fund, spent the last week of August in the area. During these periods, many members of the Harvard Field School were helpful companions in field work. For the opportunities afforded by the Harvard Field School, the authors desire to express their gratitude to Mr. Robert W. Sayles, donor of the Sayles Fund, and to Professor K. F. Mather, director of the Harvard Summer School. H. T. U. Smith and Edward Schmitz prepared the illustrations for publication.

ICE AS AN AGENT OF ROCK WEATHERING: A DISCUSSION

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ABSTRACT

The force exerted when water freezes in rock is discussed, and the process is shown to be more complex than it is ordinarily considered to be. The pressure may vary from 1 to 2,045 atmospheres per square inch, depending upon prevailing conditions and upon the manner in which freezing takes place. Perhaps it is closer to the lower figure more often than it is to the higher one. Forces exerted as a result of temperature effects upon the ice, once it has been formed in rock, are not very great.

INTRODUCTION

Geology, as a science, is in its adolescence. It still possesses the superficial, descriptive, qualitative characteristics of youth, but occasionally it exhibits the quantitative aspects of maturity. Accurate numerical data on geologic processes are difficult to obtain because a large number of factors usually are involved. For this reason, even data which have appeared in the literature repeatedly should be scrutinized before conclusions are based upon them. There is a normal tendency to accept figures which have been repeated by writer after writer, but repetition is not synonymous with validity, as will be illustrated by a specific example.

Ice as an agent of rock weathering is discussed so tritely in geologic textbooks that the reader is unaware of how poorly its action is understood. The broad scope of these books makes a brief treatment of the subject necessary. They usually emphasize the effects produced by the freezing process, such as (1) heaving of soil; (2) ascent of boulders through soil; (3) disruption and disintegration of rock; (4) formation of steep cliffs and serrate divides; (5) formation of scree, talus, and soil; (6) induration or cementation of soil by the ice, temporarily rendering the soil less pervious and less subject to transportation by wind and by water; (7) preparation of material for further weathering and for transportation. They point out less adequately the factors which are necessary or favorable to the process,

such as (1) temperature at or below 0° C.; (2) abundant moisture to saturate the rock; (3) complete incasement of the water within the rock; (4) rock having fine pores, high porosity, high permeability, favorable structure (joints, bedding planes on edge, schistosity), and low strength, either inherent or induced; (5) rapid freezing; (6) repeated freezing. Few, if any, textbooks discuss the mechanics or basic principles involved when water changes to ice in a rock, most of them attempting to convince the reader of the efficacy of the process by including a pseudo-quantitative statement of the pressure which is developed as the water freezes.

PRESSURE DEVELOPED WHEN ICE FORMS

What force is exerted by water freezing in a rock? Textbooks give conflicting answers to this question. For example, Allison writes: "Although the pressure is much less than the crushing strength of most hard rocks, it is sufficient to disrupt soft rocks or rocks weakened by cracks or by partial decay."[1] Bowles, in opposition, says: "In freezing, water expands about one tenth of its volume, and pressure exerted by this expanding force is so great that no stone is strong enough to withstand it."[2] Perusal of many American and a few English textbooks on geology indicates that most writers seem to follow Geikie, who states: "At 30° Fahr. the pressure must amount to 146 atmospheres, or the weight of a column of ice a mile high, or 138 tons on the square foot. Consequently when water freezes at a lower temperature, its pressure on the walls of the enclosing cavity must exceed 138 tons on the square foot."[3] This is approximately one ton, or 2,000 pounds per square inch, which is another figure often quoted. Geikie, like most writers on the subject, does not indicate how he calculated this quantity. Branson and Tarr,[4] very recent authors in the field, give the pressure as 150 to 200 pounds per

[1] W. H. Emmons, G. A. Thiel, C. R. Stauffer, and I. S. Allison, *Geology* (New York: McGraw-Hill Book Co., 1932), p. 37.

[2] Oliver Bowles, *The Stone Industries* (New York: McGraw-Hill Book Co., 1934), p. 354.

[3] Archibald Geikie, *Text-Book of Geology* (3d ed., New York: Macmillan Co., 1893), p. 414.

[4] E. B. Branson and W. A. Tarr, *Introduction to Geology* (New York: McGraw-Hill Book Co., 1935), p. 60.

square inch, while Scott explains: "The pressure exerted by solidifying water increases as the temperature is lowered, reaching a theoretical maximum at −8° F. (−22° C.), when it rises to nearly 34,000 pounds per square inch."[5] Here are three different answers to the question, and they vary from 150 to 34,000 pounds per square inch. Which one shall we accept?

The correct answer cannot be obtained from the consensus of opinion of a number of textbook writers, for most of them repeat Geikie's figure without qualification. Geikie specified a definite set of conditions under which water exerts the pressure as stated by him, but most authors using Geikie's data have overlooked this fact. Correspondence between several of them and the present writer suggests that one quotes the other without consulting the original reference much less stopping to verify the figure. There is a real need for accurate numerical data in geology.

Most writers err in giving the reader the impression that the freezing of water in rock is a very simple process. They imply that the force developed can be calculated from the volume change of 9.05 per cent that takes place as water changes to ice at 0° C. under atmospheric pressure and from the compressibility of ice. Data on ice are not as abundant as might be expected for so common a substance. The only compressibility figure for ordinary ice (Ice I) given in the *International Critical Tables* is that for ice at −7° C. under 300 megabaryes pressure. Under these conditions, the fractional volume change is 12×10^{-6} per megabarye.[6] Using this as the best figure available, one obtains 7.58×10^{3} megabaryes per square centimeter, or approximately 110,000 pounds per square inch, as the pressure developed when water freezes. Common experience with the behavior of ice under pressure causes one to doubt this figure, for ice, unlike most solids, melts when compressed. The process is more complex than has been assumed.

More detailed investigation of the problem brings up the work of

[5] W. B. Scott, *An Introduction to Geology* (3d ed.; New York: Macmillan Co., 1932), Vol. I, p. 218.

[6] L. H. Adams, "Compressibility of Crystalline Compounds, Minerals and Rocks," *International Critical Tables*, Vol. III (1928), p. 50.

Bridgman,[7] who, as well as Tammann, has studied the pressure-temperature relationships in the one component system containing only H_2O. The phase diagram worked out by Bridgman shows that as the temperature of a closed system filled with water is lowered, ice begins to form at 0° C. under atmospheric pressure, but only a small amount of ice can form at this temperature, for the transformation taking place sets up a pressure which depresses the freezing point. This transformation also is accompanied by the liberation of heat, 79.7 calories per gram of water changed to ice. As this heat is removed and as the temperature falls, freezing takes place under gradually increasing pressure, Ice I and water being in equilibrium until −22° C. is reached. At this temperature the pressure attains a maximum value, 2,115 kilograms per square centimeter or 2,045 atmospheres per square inch, and a new form of ice (Ice III) appears. A triple point exists at which Ice I, Ice III, and water are in mutual equilibrium. According to Bridgman,[8] Ice III is 20 per cent more dense than Ice I and 3 per cent denser than water under the existing conditions; hence any further abstraction of heat produces first a conversion of all of the water to ice without change in pressure on the system and then a conversion of Ice I into Ice III with an accompanying decrease in pressure. From this it is evident that the highest pressure that can be exerted when water freezes in rock is 2,045 atmospheres, or approximately 30,000 pounds per square inch, which is close to the figure given by Scott.

The development of this theoretical maximum pressure is dependent upon the simultaneous fulfilment of four necessary conditions: (1) the water must be confined completely so that it cannot migrate from one part of the rock to another as the pressure increases; (2) the system must consist of only one component, water, which initially is present only in the liquid phase; (3) the temperature must be −22° C.; (4) the rock confining the water must be strong enough to withstand the pressure.

The probability that all these conditions will be fulfilled during the

[7] P. W. Bridgman, "Water, in the Liquid and Five Solid Forms, under Pressure," *Proc. Amer. Acad. Arts and Sci.*, Vol. XLVII (1912), pp. 441–558.

[8] P. W. Bridgman, "High Pressures and Five Kinds of Ice," *Jour. Franklin Inst.*, Vol. CLXXVII (1914), pp. 330.

natural freezing of water in rock is not very great. Bridgman found that it is extremely difficult to keep liquids confined under high pressure; therefore it is unlikely that water will remain confined in a rock within the frost zone and exert the maximum pressure. Taber[9] has shown that when water freezes in soils, the system usually acts like an open one. The expansion which takes place as ice is formed is compensated by the expulsion of water in the voids below the frost zone and by the expulsion or compression of air in the voids in which the freezing is taking place. According to Taber, the heaving action is due to the building-up of ice layers within the soil through the growth of ice crystals. He has shown experimentally that in well-consolidated clays the surface uplift is equal to the combined thickness of the ice layers. The principles and factors involved are discussed by Taber, so there is no need to repeat them here; but they should be summarized in the textbooks on geology, especially in those for the more advanced student, and in those which emphasize the engineering and agricultural phases of the subject. Not only may the general heaving of soil be explained in the light of Taber's experiments, but the upward migration of boulders through the soil may be due to the growth of ice columns beneath them. The isothermal surface at which freezing takes place is depressed in the vicinity of large rock masses because (1) the heat capacity of rock saturated with moisture is less than that for soil saturated with moisture, (2) the heat conductivity of the former is greater than that of the latter, (3) there is less total water in the rock mass than in the adjacent soil so that there is less total heat liberated as the water is converted into ice. For a time, therefore, the lower surfaces of boulders are at the freezing point while the adjacent soil is not. This condition is conducive to the development of ice crystals beneath the boulders. As the ice layer grows, the boulders are pushed upward, provided the pressure is less than that which will cause the ice to melt at the existing temperature.

As in soil, water may freeze in consolidated rock as it would in an open system. In this case rock disintegration is nil, for the only force is that exerted to push back the unfrozen water and to expel or com-

[9] Stephen Taber, "Frost Heaving," *Jour. Geol.*, Vol. XXXVII (1929), pp. 428–61; "The Mechanics of Frost Heaving," *ibid.*, Vol. XXXVIII (1930), pp. 303–17.

press the included air. However, freezing of water in consolidated rock is not entirely analogous to that taking place in soil, for the grains in the common rocks are larger than those in soils, and they are bound tightly together instead of being loose and free to move as they are in soil. Taber has shown that ice segregation does not occur if the average soil particle is greater than two microns. The average mineral particle is much coarser than this. Taber also found that fine-grained materials when consolidated usually offer considerable resistance to fracture at low temperatures even when saturated with water, especially if the heat is conducted away in only one direction.

Under certain conditions water must freeze in rock as it would in a closed system, for the disruptive effect of the process is well known and it has been employed to quarry stone. However, with the possible exception of water included in glasses and in minerals, there probably is no water that is trapped completely within rock. The mere fact that water has seeped into an open space in the rock indicates that the system is not closed. It may become closed by cementation of the cavity walls after the water has entered. The ice itself forming in the rock may serve as an effective seal. One would expect crystallization to begin on the cavity wall, thereby trapping water within the cavity. The system also may be converted into a closed one by rapid freezing.

Water has a marked tendency to cool below the freezing point. Once the temperature has fallen below this point, it may drop 5° or 6° before crystallization takes place. Solidification may then progress with such rapidity that there is little or no time for the resulting expansion to be compensated by expulsion of some of the contents of the system, and considerable pressure may develop. These explanations must apply when automobile radiators and pipes burst as water freezes in them. Both of these systems are open, although under some hydrostatic head. There is abundant opportunity for water to move so long as freezing does not occur, but when it does, migration of the water is restricted, and a pressure builds up which often is sufficient to burst the container. Freezing of water in rock must be very similar to this, and, in rock, migration of water must

be even more limited, for the passages are more irregular and smaller, often of capillary size, so that friction and surface tension become more effective in retarding the movement of the water. Moreover, the small openings may be sealed more rapidly and more effectively by the ice which forms in them. Unlike the bursting of pipes and radiators, the disintegration of rock is not the result of a single frost but is the ultimate effect of repeated freezing and thawing, the ice acting as a wedge to widen the fissure a little more each time a frost occurs.

In order that the maximum pressure may be attained, not only must the freezing take place in a closed system but it also must involve just one component, H_2O. Ordinarily this will not occur, for either free air will be present in the voids in the rock along with the water or air dissolved in the water will be forced out of solution when the ice forms. The translucency of natural ice due to the myriad of included air bubbles is well known. The presence of the air changes the system to a multi-component one and adds a gas phase which is capable of considerable compression. Once air gets into the voids of a rock, it is not easily displaced, and the rock is not likely to become saturated very quickly. For this reason, stone which has been quarried and allowed to dry out or "season" for several weeks before the first frost is not as likely to be damaged as freshly quarried stone. Under natural conditions, therefore, the maximum pressure which water is capable of developing on freezing is not likely to be attained because the openings in the rock will not be filled with water.

The temperature of $-22°$ C. which is requisite for the production of the maximum pressure is not excessively low. This is only $-7.6°$ F., which is reached quite commonly in temperate climates even at low altitudes.

The final condition that is necessary to the attainment of the maximum pressure as water freezes in rock is that the rock must be strong enough to withstand it. Some rocks can support compressive stresses of more than 30,000 pounds per square inch, but Bridgman has pointed out that the container in which water freezes is not under compression but under tension. Rocks are comparatively weak under tensional stress, the maximum tensile strength recorded for any

rock being 250 kilograms per square centimeter,[10] or approximately 3,500 pounds per square inch. Tests of the strength of rock usually are conducted on sound, dry material; therefore this figure probably is much in excess of the tensile strength of rock within the frost zone. Although ice forming in rock may serve to cement it and thereby increase its strength, the action of the weathering agents will weaken it. Moreover, in his high-pressure work, Bridgman discovered that steel cylinders can withstand internal pressures of more than 100 per cent in excess of the tensile strength of the steel. This he attributed to the ability of the steel to stretch on the inside of the cylinder, thereby transmitting part of the stress to the outside. Rocks probably do not have the ability to stretch as steel does, but even if they do their tensile strengths would not be increased sufficiently to enable them to withstand a pressure of 30,000 pounds per square inch. It is quite evident that the four conditions necessary for the development of the maximum pressure which water can exert as it freezes in rock are not fulfilled in nature.

PRESSURE DEVELOPED AFTER ICE FORMS

A discussion of ice as a weathering agent would not be complete without consideration of the behavior of ice after it has been formed in rock. Although a great expansion takes place when water turns into ice under normal conditions, a drop in temperature below the freezing point causes ice to contract. If the ice has been formed under pressure greater than atmospheric, this contraction will tend to relieve some, if not all, of the stress. If the ice has been formed under pressure close to atmospheric, contraction will tend to cause the ice to pull away from the wall of the confining crack or cavity. Strains will be set up within the ice and between the ice and the wall rock, for the coefficient of linear expansion of ice is about ten times as great as that of rock. For a temperature range from 0° C. to 100 C.,° the largest coefficient for rock is 12×10^{-6}, as recorded by Kessler.[11] For temperature ranges below 0° C. this coefficient would be smaller.

[10] D. W. Kessler, "Building Stones," *International Critical Tables*, Vol. II (1927), p. 49.

[11] *Ibid.*, p. 54.

The coefficient of linear expansion of ice, according to Andrews,[12] is 7×10^{-5} between 0° C. and −8.8° C. Under the differential contraction indicated by these figures, there will be a tendency for the ice to crack and to pull away from the cavity wall, perhaps to pull loose rock particles with it. Where large masses of ice are involved, as ice forming and contracting on a lake, these effects are realized; but it seems that such would not happen when ice contracts in rock, because the small amount of ice involved and the small temperature drop will produce a very small total contraction, and there will be a tendency for the ice to adhere either to the cavity wall itself or to the ice in the pores in the wall rock.

If ice in a cavity should crack or pull away from the confining wall as the temperature continues to drop below the freezing point, the space may be filled with more ice acquired either by direct condensation of water molecules from the vapor phase or from a thin capillary film extending below the frost zone. According to Taber,[13] water in such a film does not freeze readily. The ice which now fills the cavity at low temperature will more than fill it as the temperature rises. The ice, expanding more rapidly than the inclosing rock, will tend to exert a pressure against the cavity wall, but for any temperature this pressure cannot be greater than a fixed maximum, as can be ascertained from the pressure-temperature diagram for the system, H_2O. Pressures greater than this would cause the ice to melt, thereby decreasing the volume and restoring the equilibrium at the existing temperature. Ice on expanding need not exert any force against the confining rock. It may simply do work against the atmosphere, if the sytem is open. Ice push is most effective when large masses of ice are heated through a considerable temperature range. Rock can be disrupted or moved only when the pressure required to accomplish this is less than that which will induce melting at the prevailing temperature. Thermal expansion and contraction of ice in consolidated rocks is probably no more effective as a weathering process than is the expansion and contraction of other rock constituents.

[12] Thomas Andrews, "Observations on Pure Ice and Snow," *Proc. Royal Soc. London*, Vol. XL (1886), p. 548.

[13] *Op. cit.*

SUMMARY AND CONCLUSIONS

A discussion of ice as an agent of rock weathering has shown that the force exerted when water freezes in rock cannot be stated as a definite value unless the conditions under which the process takes place are also stated. This force cannot exceed a theoretical maximum of 2,045 atmospheres, or approximately 30,000 pounds per square inch, and it is highly improbable that this pressure is ever attained. The conditions necessary for the production of this pressure are (1) a closed system, (2) a one-component system filled with water, (3) a temperature of $-22°$ C., and (4) a container capable of withstanding the pressure. Only the temperature requirement is commonly fulfilled in nature. Water is seldom confined completely by rock; air usually is present in the voids along with the water; and there is no rock, especially rock in the zone of weathering, that can resist a tensional stress of 30,000 pounds per square inch. The maximum strength probably does not exceed one tenth of this value.

The force exerted by ice after it has crystallized in rock is not important in the normal process of weathering. Although the coefficient of linear expansion of ice is ten times that for most rocks, the forces of expansion and contraction of ice in consolidated rock are not very great because the temperature changes and the amount of ice involved are small. Moreover, the pressure never can exceed that which causes ice to melt at the prevailing temperature.

The purpose of this paper will be accomplished if it convinces the reader that weathering of rocks by ice is not a simple process but a complex one, the efficiency of which is controlled by four basic factors.

A NEW SPECIES OF PHYTOSAUR FROM BIG SPRING, TEXAS

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ABSTRACT

The phylogenetic relationships of the Phytosauria are discussed, and a new species of *Angistorhinus* is described from Big Spring, Texas.

The phytosaurs were a group of reptiles superficially resembling the crocodiles and gavials in bodily form and probably in habits and habitat, being the ecological predecessors of these forms.[1] They differ, however, in the structure of the snout, the open palate, the presence of antorbital fenestrae, distinctive pectoral girdle and pelvis, and the position of the external nares on top of the head, as compared to the crocodiles, which have the nares at the tip of the rostrum.

The modern view places the Crocodilia and Phytosauria in different branches of a common stock having a hypothetically short-nosed ancestor. Starting with this short-nosed form, the maxillaries grew into a long rostrum to produce the Crocodilia, while the Phytosauria developed the long rostrum from the anterior extension of the premaxillaries.[2] In the crocodiles the growth of the maxillaries was accompanied by a forward migration of the premaxillaries and external nares, whereas in the phytosaurs the forward growth of the premaxillaries was accompanied by no forward migration of the external nares or maxillaries.

If the Cetacea can be used as an analogy, they illustrate the phytosaurian development in some detail. The cetaceans developed the long rostrum and high nares as a factor in their adaptation to aquatic life,[3] and this probably is the reason for the similar development in the phytosaurs. Dr. Case states that the growth of the whale ros-

[1] A. S. Romer, *Vertebrate Paleontology* (Chicago: University of Chicago Press, 1933), p. 169.

[2] E. C. Case, personal communication, 1935.

[3] R. Kellogg, "The History of the Whales," *Quart. Rev. Biol.*, Vol. III (1928), p. 29.

trum was a reaction to pressure.[4] This same agent is not suggested for the phytosaurs, but the analogy can still be used. If the evidence of the whales, of which a fairly complete series is known, is accepted, we find that the external nares remained high on the cranium when the maxillaries and premaxillaries began to grow forward.[5]

The external nares are more or less even with the posterior margin of the palate in the phytosaurs and whales. Ordinarily, the nares are completely anterior to the palate, a condition probably present in the ancestral form of these two groups. From this point of view the nares may well have migrated backward to their final position. This appears to be true of the whales, the maxillaries having been thrust back and over the frontals and parietals, telescoping the latter. This telescoping evidence of backward movement of the nares is not evident among the Phytosauria, the cranial bones being in their normal position. From this negative evidence it seems that either the nares have moved back or the palate forward. Probably both of these movements took place simultaneously. Because it has been postulated that the external nares have migrated posteriorly, the species of Phytosauria having the nares farthest forward have been classed as the most primitive.[6]

The Phytosauria have also been classed by Camp[7] with the dinosaurs, especially the Saurischia. Peculiarities of the cranium of the Phytosauria show much closer affinities with the Saurischia than with the Crocodilia.

Considering the fact that the Phytosauria are confined to the Triassic, they show a large range of structural differences. This points to their value as aids in stratigraphic correlation when the faunas will have become more thoroughly known.

In October of 1931 Roy Lamb of Big Spring, Texas, notified Dr. C. E. Decker, of the University of Oklahoma, of the discovery of what promised to be a complete phytosaur on his ranch situated 25 miles south and east of Big Spring. In November the senior author, accompanied by L. I. Price, went down to collect the specimen.

[4] Personal communication, 1935.

[5] Kellogg, *op. cit.*, pp. 29–76, 174–208.

[6] C. L. Camp, "A Study of the Phytosaurs," *Univ. Calif. Mem.*, Vol. X (1930), p. 148.

[7] *Ibid.*, p. 152.

Five articulated vertebrae were found exposed on a small knoll. At each end the bone extended into the ground. Excavation, however, revealed only four more vertebrae, the skull (the mandible was not present), some rib fragments, and about 2 square feet of dermal scutes. A few small bones and teeth were found at several other places in the immediate vicinity.

All of the bones were covered with a layer of limonite about $\frac{1}{16}$ inch thick, together with patches of gypsum somewhat greater in thickness. The material in which the bones were deposited was a dark, brownish-red clay, marked with splotches of gray and breaking in very irregular pieces.

In the faces of the bluffs in this area is a conglomeratic layer which is believed to be the bottom of the Trujillo formation. If the identification be correct, this phytosaur specimen came from the upper part of the Tecovas or Lower Dockum, which is of Upper Triassic age. The only fossils heretofore reported from the Tecovas are pieces of wood, all of the animal remains being described from the Trujillo.[8] More detailed field study and correlation will be necessary, however, before the beds in which this phytosaur was found can be assigned definitely to Tecovas age.

As noted by Dr. A. O. Weese,[9] among the vertebrates the skull is the most constant specific character. Any prominent inheritable skull change in an animal is considered by zoölogists to justify establishing a new species. The skull described in this paper seems to be set apart from previously described forms, particularly by the great length and height of the skull, as well as the large angle which the planes of the orbits make with the horizontal.

On the basis of the primitive characteristics which it possesses, it should be classed with Camp's Angistorhininae. These primitive characters as outlined by Camp[10] are as follows: (1) supratemporal fenestrae normal; (2) nares subcircular; (3) narial septum thin; (4) nares anterior (to antorbital fenestrae); (5) squamosal short; (6) no enlarged rostral crests; (7) auditory meatus a notch.

Because the following non-primitive characteristics are present, it

[8] E. H. Sellards *et al.*, *Univ. Tex. Bull. 3232* (1933), p. 244.

[9] A. O. Weese, personal communication, 1935.

[10] C. L. Camp, *op. cit.*, p. 148.

most closely resembles the genus *Angistorhinus:* (1) two types of teeth (cone-shaped and cuneiform with serrated edges); (2) post-temporal openings large; (3) pineal foramen absent; (4) skull elevated.

The new species also possesses another non-primitive characteristic which is lacking in *Angistorhinus*. This is the outward-directed orbits, which appear to be better developed than in any previously described species.

It also resembles *Promystriosuchus* in many features, especially in the possession of a large skull with a long rostrum, which, as Camp notes, *Angistorhinus* lacks. However, the species *A. grandis* Mehl is one of the larger phytosaurs, being some 977 mm. in skull length.

The genus *Angistorhinus* has been carefully defined by Case,[11] who gives the more distinctive characteristics as follows: (1) post-temporal arcade at the level of top of skull; (2) antorbital vacuity with the anterior end even with the anterior end of the external nares; (3) length of the prenarial portion of the skull to the post-narial portion as 5 is to 3; (4) squamosals extended backward and with a strong descending hook; (5) orbits look upward more than outward.

The individual here described deviates from the foregoing type by having the anterior end of the antorbital vacuity slightly posterior to the anterior margin of the external nares, the length of prenarial skull to postnarial as 2 is to 1, and the orbits looking decidedly outward.

In the light of these comparisons and in spite of the discrepancies, this new species is placed in the genus *Angistorhinus*, the writers considering the existing differences to be of doubtful generic value.

ANGISTORHINUS ALTICEPHALUS, N. SP.

GENERAL FEATURES

The skull is elongate with a long, slender snout (Fig. 1). The nares are situated in a prominence at the posterior end of the rostrum. The skull is high and among the longest of described phyto-

[11] E. C. Case, "New Reptiles and Stegocephalians from the Upper Triassic of Western Texas," *Carnegie Inst. Wash. Bull. 321* (1922), p. 69.

saurs, with a length of 1,220 mm. The squamosals extend beyond the quadrates and are produced in downward-hooking processes. The supratemporal vacuities are closed by a stout parieto-squamosal arch lying in the plane of the top of the skull. The top of the skull is distinctly flat, a sharp angle existing between the top and sides of the skull. The fenestrae are all very large. The orbital edge of the pre-

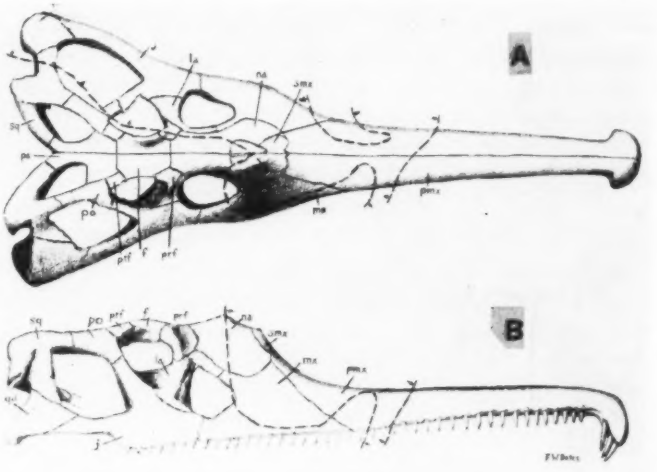

FIG. 1.—*Angistorhinus alticephalus* Stovall and Wharton, n. sp., ×1/8. *A*, dorsal view of the skull; *B*, lateral view of the skull. Dashed lines indicate borders of restored areas. Arrows point to restored portion.

frontal and frontal bones is distinctly produced into a ridge which stands up and out, over the orbit. The orbits are on the sides of the skull and look outward much more than up.

DORSAL SKULL FENESTRAE

The external nares are elevated on a prominence at the posterior end of the rostrum above the ventral edge of the maxillaries. The nares are separated by a septum which is deep set, giving the nares what appears to be a single opening on the surface. They are 780 mm. from the tip of the rostrum, with the anterior border slightly in

front of the anterior border of the antorbital fenestrae. The position of the septomaxillaries is problematical.

The antorbitals are large, subtriangular in outline, acute anteriorly, and rounded posteriorly. They measure 100 mm. in their greatest diameter, which is with the length of the skull. There is a sudden thinning of the bone into an iris-like process at the postero-ventral margin.

The large orbits are nearly circular in shape. Along the ventral and anterior margins the bone is thin and sclerotic, giving the orbit a triangular outline. There is a well-developed prominence over the dorsal margin of the orbit formed by the frontal and prefrontal bones. The orbits are 90 mm. in their largest diameter, which is parallel with the largest diameter of the antorbital fenestrae. The orbits are directed outward and slightly upward.

The post-temporal fenestrae are trapizoidal in shape with straight sides. The longer diagonal is 160 mm., while the shorter diagonal is 120 mm. There is a thinning of the bone at the antero-ventral angle.

The supratemporal vacuities are triangular in shape with an anterior-posterior length of 90 mm. and a width of 50 mm. They are closed behind by a rugose parieto-squamosal arch which lies in the plane of the top of the skull. No pineal foramen is present.

SEPARATE DORSAL SKULL BONES

The premaxillaries are produced into a long, rather slender snout, strongly hooked downward at the anterior end. They taper from the skull proper with a width of 140 mm. at a distance of 140 mm. in front of the nares to 73 mm. at the tip. The anterior tip of the rostrum is expanded to 108 mm. and is bent down 56 mm. below the ventral surface of the rostrum.

The maxillaries lie just posterior to the premaxillaries and ventral to the nasals. The suture between the maxillaries and premaxillaries was not determinable.

The external nares are completely surrounded by the nasals except on the extreme anterior margin. Here a pair of questionable septomaxillaries separate the external nares from the premaxillaries (Fig. 1).

The nasals extend from the external nares back over the antorbital fenestrae, being separated from these openings anteriorly by the maxillaries and posteriorly by the lacrimals. They terminate just in front of the anterior margin of the orbits at a distance of 76 mm. posterior to the nares.

The lacrimals form the antero-ventral borders of the orbits, as well as most of the posterior and dorsal margins of the antorbital vacuities.

The prefrontals are small, semirectangular in outline, and form the antero-dorsal corners of the orbits.

The centro-dorsal borders of the orbits are formed by the frontals, which unite to form the top central portion of the skull. These bones are 53 mm. along the median union and are set off by deeply entrenched sutures.

Just posterior to the frontals is another pair of small subrectangular bones, the postfrontals. These are bounded on the inner side from front to back by the frontals, parietals, and postorbitals.

The parietals form the posterior part of the top of the skull and are 108 mm. along the median line of the dorsal skull surface. They form the anterior and inner margins of the supratemporal fenestrae. The anterior end of the parietals is quite blunt, the suture traversing the skull at right angles to the median skull suture. This is markedly different from *Angistorhinus grandis* Mehl,[12] which has distinctly anteriorly pointed parietals, each half of the suture crossing the skull at 45°, forming a 90° angle which the median skull suture bisects.

Most of the outer margins of the supratemporal fenestrae are formed by the postorbitals. These bones extend from the posterior edge of the postfrontals 65 mm. back to the squamosals. The anterior portions of the inner margins border on the parietals, while the outer margins are extended downward, meeting the jugals and with these latter bones forming the bars between the orbits and the posttemporal fenestrae.

The remainder of the supratemporal fenestrae are outlined by the squamosals, which also form a large part of the postero-dorsal corner

[12] M. G. Mehl, "The Phytosauria of the Trias," *Jour. Geol.*, Vol. XXIII (1915), p. 130.

of the post-temporal fenestrae. The posterior ends of the squamosals are produced into descending hooklike processes extending more than 80 mm. below the top of the skull. The extreme posterior ends of the squamosals have a distinct angle where the descending hook begins as compared to *Angistorhinus grandis* Mehl,[13] which is more or less rounded, curving gradually into the hook.

The jugals form the ventral marginsof the skull from the antorbital vacuities to the posterior margins of the lateral temporal fenestrae. They also extend dorsally to meet the postorbitals, helping to form the bars between the orbits and post-temporal fenestrae.

The quadrates form the postero-ventral angles of the skull. They meet the jugals under the post-temporal fenestrae and the squamosals at the postero-ventral corner of these same fenestrae. The posterior angles of the quadrates carry rounded, barlike processes used for articulation with the mandible. These bones are the quadrate-jugals of Mehl.[14] Comparing these bones to the homologous bones of *Angistorhinus grandis* Mehl, it is seen that those of *A. alticephalus*, n. sp. (Fig. 1) are much larger, extending posteriorly until they are even with the posterior extremities of the squamosals. The sutures vary but slightly in the two species.

VERTEBRAE

The nine vertebrae found were all dorsals. They have been slightly distorted laterally, but the essential features and relative measurements are fairly well represented (Fig. 2).

The measurements compare with those of *Angistorhinus gracilis* Mehl but are in nearly every case slightly larger. The centrum is laterally compressed, measuring about 30 mm. thick in the center.[15] The anterior and posterior ends are oval in outline, the anterior end usually being about 5 mm. smaller in diameter. The transverse diameter across the face of the centrum is 63 mm., while the vertical diameter is 75 mm. on the anterior face. The neural arch has been greatly compressed during burial, the neural canal being entirely closed in some cases.

The neural spine is 120 mm. high, measured from the top of the

[13] *Ibid.* [14] *Ibid.*, p. 135. [15] *Ibid.*, p. 153.

centrum. The top end of the neural spine is expanded into a broad, concave, distal terminus measuring 50 mm. in diameter across and 60 mm. parallel with the neural canal. The rugose transverse processes have been lost in each case, leaving broken proximal ends tri-

FIG. 2.—Dorsal vertebrae of *Angistorhinus alticephalus*, n. sp. *A*, anterior view; *B*, lateral view.

angular in cross section with one vertex pointing downward. These proximal ends are 30 mm. across. The knoblike heads upon which the ribs articulated are 15 mm. in diameter. The zygopophyses are broad, flat, and oval in outline, with a long diameter of about 35 mm. The anterior zygopophyses are directed upward and inward at an angle of about 30° with the horizontal, and measure about 40 mm. between the centers of the faces. The total height of the vertebrae is 205 mm.

ANGISTORHINUS GRACILIS COMPARED TO *A. ALTICEPHALUS*

	*A. gracilis**	*A. alticephalus*
Length of centrum	45 mm.	70 mm.
Height of centrum	54 mm.	85 mm.
Total height of vertebrae	150 mm. (at least)	205 mm.
Transverse diameter of anterior centrum face	69 mm.	65 mm.
Diameter of centrum at center	29 mm.	29 mm.
Length of skull	950 mm.	1,220 mm.

* All measurements of vertebrae of *Angistorhinus alticephalus*, n. sp., are average.

A further comparison of *Angistorhinus gracilis* and *A. alticephalus* tends to show that the two species are distinct.

A. gracilis	*A. alticephalus*
Terminal expansion of rostrum takes place suddenly and keeps the same width to the tip.	Terminal expansion takes place gradually, enlarging toward the tip.
Rostrum 56 mm. wide at a point 140 mm. in front of antorbital vacuities.	Rostrum 180 mm. in width at a point 140 mm. in front of antorbital vacuities.
Greatest length of skull 985 mm.	Greatest length of skull 1,220 mm.
Length from anterior border of orbits to tip of rostrum 750 mm.	Length from anterior border of orbits to tip of rostrum 970 mm.
Interorbital width 68 mm.	Interorbital width 85 mm.

ACKNOWLEDGMENTS.—Thanks are due to Dr. E. C. Case for many valuable suggestions while the paper was still in manuscript. And to Mr. F. W. Bates we wish to express our deep appreciation for the excellent drawings.

TRAVERTINE-DEPOSITING WATERS NEAR LEXINGTON, VIRGINIA

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ABSTRACT

The cascading, travertine-depositing part of Wilson Falls Creek is supersaturated with respect to $CaCO_3$ throughout the year, owing to the supersaturation of the feeder springs. The creek loses $CaCO_3$ when its temperature is above 8° C. The losses increase with rising temperature, but at all times adjustment is very slow. The rate of adjustment is accelerated by rising temperature, aëration, and close contact of the water with calcite.

INTRODUCTION

During the interim from May, 1934, to May, 1935, monthly tests were made on the lower, cascading, travertine-depositing portion of Wilson Falls Creek and on one of its large feeder springs, which issues near the head of the rapids.[1] These waters drain a limestone and dolomite area 18 miles northeast of Lexington, Virginia. The creek was tested (1) immediately below the feeder spring; (2) at the top of Cypress Falls; (3) the bottom of Cypress Falls; and (4) the mouth of the creek. The first and last stations are about a mile apart, and the drop at the falls is 130 feet.

THE TEMPERATURE AND CHEMICAL COMPOSITION OF SPRING AND CREEK

In the summary of the temperature and chemical characteristics of the spring and stream given in Table I, the annual range of the constituents is given in parts per million. The figures in column 5 show the theoretical content of $CaCO_3$ in solutions adjusted to the partial CO_2 pressure of the air and to the temperature range given in column 3. The content assigned by theory to the saturated solutions is taken from Pia's *Kohlensäure und Kalk*. For the sake of brevity, the solubility of $CaCO_3$ as used in this paper is really its solubility in various concentrations of CO_2-bearing waters.

According to Table I, all the waters are supersaturated with

[1] Edward Steidtmann, "Travertine near Lexington, Virginia," *Science*, Vol. LXXX, No. 2068 (1934), pp. 162–63.

$CaCO_3$, the amount of saturation being about equal to the difference between the figures in columns 4 and 5. The spring is more supersaturated than the stream. Doubt as to the verity of supersaturation was dispelled by aërating the waters in the presence of calcite. When this was done long enough, the $CaCO_3$ content of the waters was reduced to a constant concentration, which was about the same as that which theory has assigned to saturated solutions of the same temperature when under a partial CO_2 pressure of 0.0003 of an atmosphere, this being about that of the CO_2 in the air. Incidentally,

TABLE I

ANNUAL RANGE OF THE pH, TEMPERATURE, AND CONSTITUENTS OF THE TRAVERTINE-DEPOSITING WATERS OF WILSON FALLS CREEK AND OF A FEEDER SPRING ISSUING NEAR THE HEAD OF THE TRAVERTINE DEPOSITS, IN THE INTERIM FROM MAY, 1934, TO MAY, 1935

	pH (1)	Free CO_2 (2)	Temperature (3)	$CaCO_3$ (4)	$CaCO_3$ Saturated Solution (5)	$MgCO_3$ (6)	SO_4 (7)	Cl (8)
Feeder spring....	7.1	19	11°–15° C.	165–90	65–68	41–52	1>5	tr.
Wilson Falls Creek.....	7.6–8.2	0–3	6.5°–25° C.	123–56	55–80	37–86	1>5	tr.

the similarity between the theoretical concentration of saturated solutions and of the spring and creek waters when reduced to saturation shows that the compounds in solution other than $CaCO_3$ do not occur in sufficient concentration to affect its solubility to any extent.

When the temperature of the stream was below 8° C., the $CaCO_3$ content of the stream was the same, or nearly the same, at the four sampling stations, showing that little or no $CaCO_3$ was falling out. At higher temperatures, the analyses showed a progressive loss of $CaCO_3$ in the downstream direction, the rate of loss increasing with the temperature. The greatest difference between the $CaCO_3$ of the first and last stations was over 20 parts per million. Rise in temperature seems to be the main factor which hastens the adjustment of the supersaturated solutions toward equilibrium with the air.

WHY SPRING AND CREEK WATERS ARE SUPERSATURATED

The supersaturation of the spring with respect to both CO_2 and $CaCO_3$ is interpreted as resulting from the high CO_2 content of soil and subsoil gases, reported as commonly ranging from 1 to 8 per cent by volume. This is from 33 to 270 times greater than the CO_2 content of the air. Therefore, it is logical to find in the Lexington area the $CaCO_3$ content of the springs to be in some direct relation to the humus cover, those in unbroken forest having about 225 parts per million, which is about three times greater than springs issuing on closely grazed, hillside pastures.

But the adjustment of the supersaturated waters to the air is amazingly slow. In view of their supersaturation, it is surprising, not that they deposit calcite, but that they deposite so little. The physical conditions demand deposition, but the response is surprisingly feeble. That this response is hastened by rise in temperature has been indicated already. Experiments outlined in this paper show that this response is also accelerated by aëration and by the presence of calcite. The importance of calcite as a catalyst was hardly foreseen.

PLANTS, OF LITTLE EFFECT ON ADJUSTMENT AT PRESENT

It is improbable that the thin film of algae which coats most of the cascades at present accelerates the stabilizing reactions of the relatively thick sheet of water which dashes over it. The algal film seems about the same throughout the year, being affected more by floods and occasionally by ice than by the seasons; but the deposition of calcite is seasonal, and its relation to a surge in plant life is not in evidence.

At an earlier day, when the creek waters flowed in thin sheets and films over travertine cliffs coated with algae and mosses, the effectiveness of plants in stabilizing the waters may have been greater than now. In the prehistoric heyday of travertine building, Wilson Falls Creek was about twenty times wider and correspondingly shallower than the present stream. The waters flowing out of primeval forests were initially even richer in $CaCO_3$ and CO_2 than the average spring of today. With the advent of cultivation the $CaCO_3$ content of the springs has declined, and their flow has become more

irregular. Floods have caused the streams to become entrenched on the travertine deposits to depths of 25 feet, instead of spreading out in thin sheets over depositional surfaces where the water was in intimate contact with air and calcite. The streams now have cut to bed rock in most places and no longer nourish a rich underflow of water in the travertine pool and channel deposits—an underflow which emerged in seeps and trickles on the travertine cliffs farther down the valley. The cascade travertine cliffs are now dry, the falls having receded about 100 feet from the cliffs, along a narrow channel. None of the primeval streams remain. Their deposits, however, are proof of their having been adjusted much more rapidly than the stream of today, but it is difficult to say whether their more rapid adjustment was directly due to the life processes of the plants or to the waters having spread in thin sheets in closer contact with air and calcite.

The physical relation of the plants to the associated travertine is well illustrated by the moss *Philonotis calcarea*, which grows best where thin sheets of water or trickles flow over the face of the cascades. Its living leaves show no calcite except on the basal parts of the plant, where it is in contact with, or in close proximity to, previously existing calcite. Cascade algae show similar traits. The deposition of new calcite on or near previously existing calcite, rather than on the green top foliage, is interpreted as a result of the catalytic effect of calcite on the rate of adjustment of supersaturated $CaCO_3$-bearing waters. Most of the deposition on cascades is a direct addition to previously deposited materials, the calcitic deposit taking place around the base of a felt of plants the surface of which usually remains uncoated. Thus, otherwise rootless plants are anchored in the walls of the cascades, where they benefit by contact with aërated, CO_2-yielding waters. The felty verdure, in turn, forms a framework for the calcitic deposit and canopies the growing crystalline mass against erosion—a unique case of co-operation between living and crystal matter.

ADJUSTMENT AIDED BY AËRATION AND BY THE CATALYTIC EFFECT OF CALCITE

That aëration and intimate contact with calcite influence the rate of adjustment of supersaturated $CaCO_3$ solutions is indicated by the

following experiments. A milky suspension of calcite was held up by vigorous aëration in a solution containing 200 parts of $CaCO_3$ per million. Aëration was carried on by the use of a suction pump. The temperature was 21°–22° C. This solution was adjusted by 28 hours of aëration. A similar passage of CO_2-free air would have changed the same volume of ordinary distilled water into a standard CO_2-free water in about 3 hours. The $CaCO_3$ content of the adjusted solution was 53 parts per million, which was nearly the same as the amount assigned by theory to a saturated solution at this temperature. Water from Wilson Falls Creek gave similar results.

Another solution containing 200 parts of $CaCO_3$ per million was aërated in the same way as in the preceding case, but it was freed from calcite by repeated filtration. After 19 days of this treatment, the solution still had 70 parts of $CaCO_3$ per million; yet additional aëration brought down no more, although the solution had about 15 parts per million in excess of saturation. The temperature was 21°–22° C. Adjustment by vigorous aëration was therefore about 19 times faster with calcite present than in its absence.

Samples of the aërated solutions, when placed over calcite in stoppered bottles without an air space above the solutions, lost considerable $CaCO_3$ after 12–24 hours of standing when the $CaCO_3$ content was above 80 parts. Free CO_2 was released in this reaction. Check solutions without calcite, when put into flasks, showed no appreciable change in the same time interval. The liberation of free CO_2 in the flasks containing calcite brings to mind the high content of free CO_2 frequently found in deep-sea waters close to calcitic deposits. It also raises the question whether the non-skeletal, calcitic muds are largely precipitated on the bottom close to previously existing calcite or whether they are more commonly thrown out at higher levels and then sink to the bottom.

Two solutions of the same volume and depth, each with 330 parts of $CaCO_3$ per million, stood in air for 10 days—one with, the other without, calcite. The one with calcite had been reduced to 61 parts of $CaCO_3$ per million; the other, to 95 parts per million. After 21 days the score stood: with calcite, 51 parts; without calcite, 57 parts. The temperature of both was 21°–22° C. Other tests showed

that adjustment was almost twice as fast with calcite as without calcite under conditions of quiet standing.

Repeated experiments showed that adjustment of the supersaturated $CaCO_3$ solution, either by standing in air or by artificial aëration, with or without the presence of calcite is a slow process. Creek waters and prepared solutions were about the same in this regard. The first hour of aëration of the creek waters in the presence of calcite usually drops about 30 parts of $CaCO_3$ per million, the bulk of the excess dropping in the first 10 hours. In the course of aëration, the pH rises to a maximum of about 8.2 and then declines to about 7.6. The spring water, being initially high in free CO_2, does not lose $CaCO_3$ in the first hour of aëration with calcite, but usually shows a slight gain. The slowness with which such solutions give up their excess in the laboratory merely parallels nature. In summer, the 130-foot drop of Wilson Falls Creek at Cypress Falls merely results in a loss of about 10 parts of $CaCO_3$ per million. On a hot day, Falling Springs north of Covington, Virginia, lost 10 parts per million in a sheer drop of 60 feet, the water clearing the cliff. In this case the temperature of the water rose 1° C. in the fall, and the waters did not come in contact with visible plants.

POSSIBILITY OF COLLOIDAL CALCITE AS A PRODUCT OF ADJUSTMENT

It may be that the adjustment of supersaturated $CaCO_3$ solutions is actually somewhat faster than it seems. If some of the calcite thrown out in the course of adjustment should be too fine grained to be retained by the filter paper, its crystalline, rather than dissolved, state would not be disclosed by the usual methods of titration. The true state would be disclosed by determining the total CO_2. If some of the $CaCO_3$ is present in the colloidal state, the total CO_2 would fall short of the amounts indicated by titration methods. In view of the alleged precipitation of colloidal $CaCO_3$ by the sea believed probable by some writers, NaCl was added to samples of creek water up to 5 per cent. This did not change the $CaCO_3$ content of the water in the slightest. Either the creek water has no colloidal calcite or else NaCl does not coagulate colloidal calcite. Incidental-

ly, it is interesting to recall that NaCl greatly increases the solubility of $CaCO_3$ in CO_2-bearing waters.

BEARING OF STUDIES ON THE INTERPRETATION OF CAVE CALCITE

The twin rôles played by the partial CO_2 pressure of the air on contact with supersaturated $CaCO_3$ solutions in the precipitation of calcite, and by calcite as a catalyzer of this reaction, raises the question whether evaporation is as important in the deposition of cave calcite as has sometimes been supposed. What is really known about either the humidity or the CO_2 content of the air in caves? How much is known about the CO_2 content of cave waters? A priori, two extreme types of limestone caves probably exist: those which have, and those which lack, an effective air circulation. In those having such a circulation, the CO_2 partial pressure would be similar to that of the air outside; and this would favor the precipitation of calcite from the supercharged solutions. Such a circulation would also raise the summer temperatures of the cave, which would also hasten adjustment. Because of the catalytic effect of calcite, adjustment should be most rapid where thin films of water flow in contact with both air and calcite. Conversely, in a stagnant cave the CO_2 content of the air should tend toward a condition of equilibrium with the highly carbonated waters of the cave, and this would inhibit the deposition of calcite.

On theoretical grounds, evaporation may have no effect on the adjustment of $CaCO_3$ solutions, except in so far as it would give rise to calcite films, which in turn would exert a catalyzing effect on adjustment. The question whether evaporation in itself has any effect on the rate of adjustment is raised nevertheless. It was found that adjustment was nearly twice as fast from an evaporating solution in an open beaker as from a beaker of the lipped type having a cover glass, from which there was practically no evaporation. Supersaturated $CaCO_3$ solutions kept in large, loosely covered cans whose lids were lifted daily for a few moments lost their excess more slowly than samples kept in the open and subject to evaporation. In these cases the more rapid adjustment of the open samples is probably not due to evaporation. Both the more rapid adjustment and more rapid evaporation probably are the result of the same condition, namely,

that in the open both escaping CO_2 and water vapor are quickly removed by convection; and thereby any rise in the partial pressure of these gases directly above the solution is prevented. The cases cited above illustrate, to some extent, the difference between adjustment in the cave with stagnant air and in the one with circulating air. The question is still raised: Does evaporation increase the rate of adjustment of supersaturated $CaCO_3$ solutions?

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WIND-FACETED STONES FROM MARLBOROUGH, NEW ZEALAND

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ABSTRACT

Owing to the surface configuration in this locality, the winds blow alternately from opposite directions, carving sand-blast facets on flat limestone fragments. The arrangement of the facets with regard to wind directions, shape and position of the fragments, and the attitude of the base is discussed, and an account given of the stages in the alteration of the shape of pebbles as the process continues. Two main types are discriminated: those with a broad base relative to the height, which tend to be reduced to flattish or broadly convex plates; and those with a relatively small base, from which more irregular forms are produced. Wind-faceted stones are classified on this basis and a terminology provided.

INTRODUCTION

The earliest description of the faceting of pebbles by a blast of wind-driven sand is that of W. T. L. Travers[1] (1870), who, rejecting the theory of native manufacture, ascribed such an origin to certain faceted pebbles at Evans Bay, Wellington, New Zealand. Since then, several observers have recorded the occurrence of similar products on different portions of the New Zealand coast.[2] The areas described are all characterized by sandhills which are subject to the action of strong winds. Nowadays, in spite of the multitudes of native artifacts which are also found in such situations, no reasonable doubt is expressed that these characteristic stones were once irregularly shaped rock fragments and that they have been faceted and worn by the abrasive action of innumerable sand grains under the influence of the winds which blow on these exposed portions of the New Zealand coast.

In spite of their frequent occurrence and early recognition, however, no systematic account of their development has yet been given;

[1] *Trans. N.Z. Inst.*, Vol. II, pp. 247–48, Pl. XVI.

[2] P. Thomson, *Trans. N.Z. Inst.*, Vol. III (1871), p. 269, from near Dunedin; W. B. Dawkins, *Quar. Jour. Geol. Soc.*, Vol. LVIII (1902), p. lxxxv, from Waitotara; E. Stowe, *Trans. N.Z. Inst.*, Vol. V (1873), pp. 105–6, from Waikato; E. W. Andrews, *ibid.*, Vol. XXVI (1893), p. 397, from Wanganui; C. A. Cotton, *The Geomorphology of New Zealand* (N.Z. Govt. Printer, 1922), p. 258, from Wellington; F. A. Bather, *Proc. Geol. Assoc.*, Vol. XVI (1900), p. 402, from Hokitika.

and as the writer enjoyed peculiarly favorable opportunities for observing the mode of formation of specimens in Marlborough, these will be taken as the basis of the present discussion.

The pebbles occur on a narrow strand plain fringing the northern extremity of the east coast of the South Island, where a narrow strip of sand beach borders high coastal cliffs. These cliffs effectually cut off any winds from the interior of the island, and cause one of the

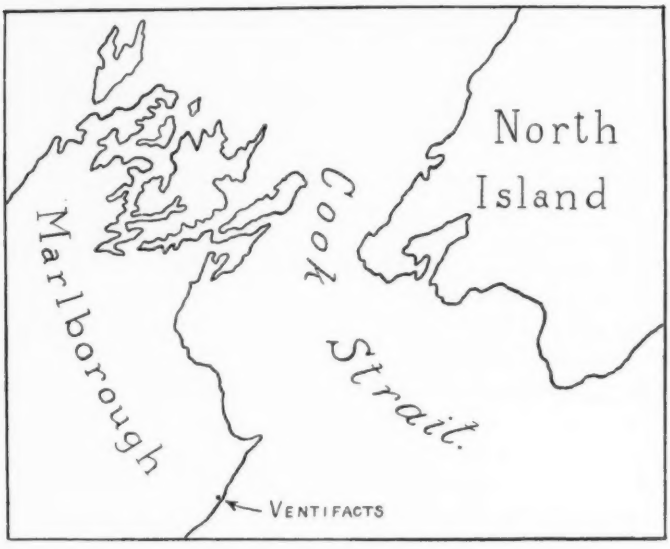

FIG. 1.—Locality map

two prevailing winds (the northwest) to assume a local north-northeasterly direction, i.e., directly down the coast parallel with the shore. The other prevailing wind is the south-southwest, which blows in the opposite direction. Only these two provide that long "fetch" and drift of sand which is an effective abrading agent. The few easterly winds which occasionally blow in from the sea lack the same intensity and are ineffectual sand-movers, as the strand plain is too narrow for any pronounced drift to occur across it. Any pebbles on the strand plain are therefore subjected to a heavy bombardment of sand grains alternately from the north-northeast or the

south-southwest. Conditions are thus similar to those originally described by Travers from Evans Bay, on the other side of Cook Strait, where, owing to the conformation of the hills, the winds blow from either a northerly or a southerly direction.

At the same point on the Marlborough coast is an outcrop of a fine-grained, flaggy, whitish limestone which supplies, on weathering, a great number of small, flat slabs ranging in thickness from $\frac{3}{4}$ inch to 2 inches. These fragments constitute the raw material from which the various forms of ventifacts are carved. The most abundant fragments are those which are triangular or rectangular in form, but more irregular shapes are not uncommon.

THE MANNER OF CUTTING

In 1910, Wade[3] demonstrated in the field that the great majority of wind-shaped stones lie with their front facets *across* the wind, and even earlier Bather[4] had noted that such observations as were then available indicated that this was true. Pushing this conception further, he expressed the opinion that each facet was cut at right angles to a separate wind direction. This is in direct opposition to the views of Heim,[5] who considered that the longer axis of the pebbles (and the facets) should lie parallel with the direction of air currents near the ground. The experience of the writer is that, though a majority of the Marlborough specimens have the directions of the facets determined by factors other than wind direction, yet many do seem to be orientated with the larger facets across, rather than parallel to, the wind. The selected examples in Figure 2, though not entirely representative, illustrate this point.

THE SIMPLEST SEQUENCE OF FORMS

In the case of a broad, rectangular fragment so placed that it lies directly across the wind, i.e., presents a side face[6] to each of the prevailing winds, two simple facets are cut, sloping upward toward each

[3] A. Wade, *Geol. Mag.*, Vol. VII (1910), p. 396.

[4] *Op. cit.*, pp. 493–94.

[5] Albrecht Heim, *Vierteljahrschr. Zuricher naturf. Ges.*, Vol. XXXII (1888), pp. 383–85.

[6] In this paper the term "face" is restricted to *original* surfaces of the rock fragment. "Facet" is applied only to those surfaces which have been cut under wind action.

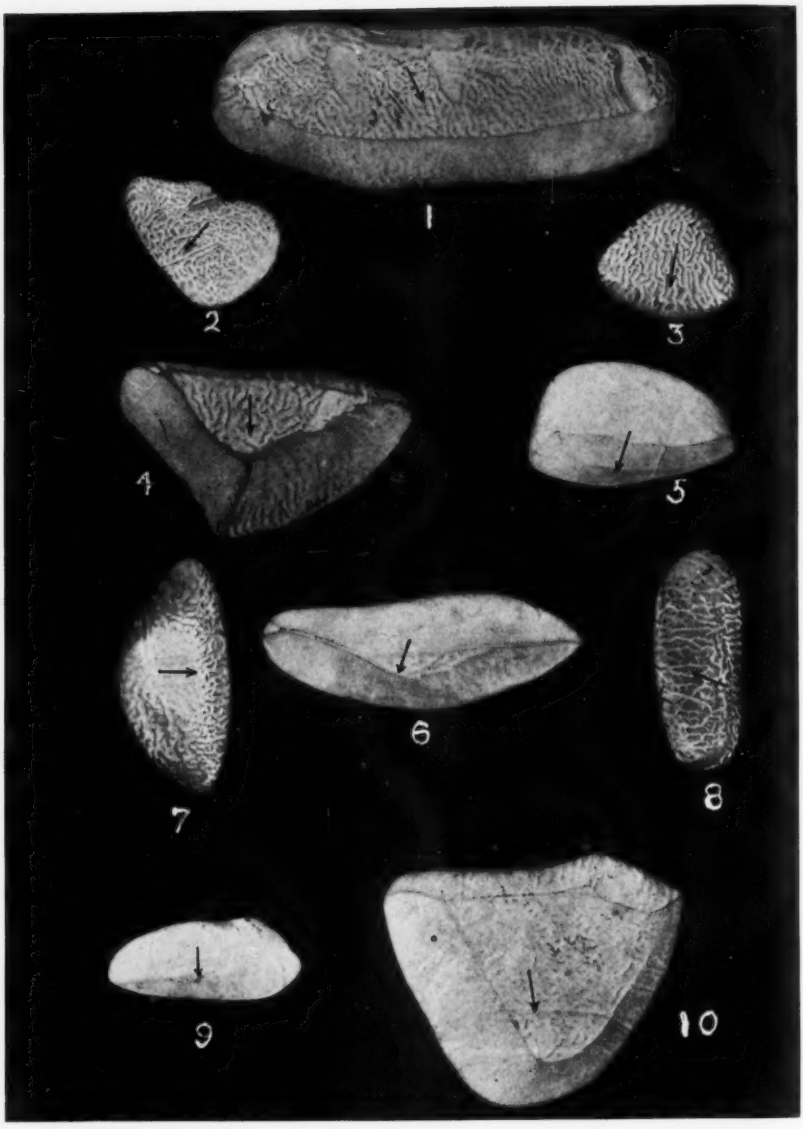

FIG. 2, ILLUSTRATIONS 1–10.—Wind-faceted stones from the strand-plain, 2 miles north of the mouth of the Ure River, Marlborough, New Zealand. (The arrows point north-northeast.)

other, the ends of the facets being slightly beveled by the blast. Under the continued influence of the sand drift the slope of these facets is gradually reduced until they reach a low angle, and in the final stage an almost flat surface is produced. At the same time the facets retreat toward each other as a whole, so that the size of the base diminishes. The whole process, however, is not uniform. At first, when the face is nearly vertical, cutting is most rapid, and the facets developed soon attain an angle of 50° or 60° from the horizontal. After this relatively stable stage is reached, reduction in the angle of slope becomes more and more retarded, and the facets retreat more and more as a whole, so that pebbles which have reached a relatively late stage are rare. The experiments of Thoulet with an artificial sandblast (quoted by J. B. Woodworth)[7] yielded similar results: "The abrasion is so much the more energetic when the rock upon which it acts is more nearly vertical with reference to the direction of the sand which strikes (horizontally), and it diminishes very rapidly in intensity directly as the inclination becomes lower than 60 degrees." Such a sequence of ideal forms must be of rare occurrence, as the conditions for its development are very restricted; but specimens formed under comparable conditions from triangular fragments show a parallel evolution and will be described in a subsequent section.

On the limestone slabs the upper, flat surfaces and the low-angle facets are also characterized by small, irregular furrows of the typical *rillenstein* appearance (Fig. 2). These irregularities, developed where wind action is at a minimum, may perhaps best be interpreted as solution effects, and open up the question of the rôle of solution in the final reduction of the upper surface of a fragment to a horizontal plane. The problem has been investigated by Bryan,[8] who described characteristic pebbles faceted by solution under arid conditions in America. In general his specimens have a high central area (the "boss") surrounded by a lower, peripheral "rim." None of the specimens collected by the writer show the typical "boss" and "rim" of the American examples, but Bryan himself remarks (p. 195) that "on some pebbles the boss is absent and on others the rim." That

[7] *Amer. Jour. Sci.*, 3rd ser., Vol. XLVII (1894), p. 69.

[8] Kirk Bryan, *Amer. Jour. Sci.*, Vol. XVIII (1929), pp. 193–208

solution effects are not absent in Marlborough is shown by a light deposit of travertine on the base of many stones. It would seem that, as wind action becomes less severe when the angle of the facets is reduced considerably, solution becomes relatively more important forming the *rillenstein* markings on the upper surface and depositing a thin film of travertine below.

THE INFLUENCE OF ORIGINAL DIRECTION

In a majority of cases a flat, rectangular slab of rock does not lie directly across the wind, presenting two faces at right angles to the prevailing wind directions, but assumes some position in which a vertical edge is more or less directed into the wind. Two faces, one on either side of the edge, are then directed obliquely toward the wind. Both of these are simultaneously trimmed back into facets by the sand blast. In connection with this oblique cutting, Thoulet noted: "Oblique incidence too (from 30° to 45°) increases the rapidity of the cutting effect, probably because the issuing particles of sand are not met and their force deadened by the rebounding ones."[9] Observation of triangular specimens from Marlborough showed that on the side where one face was presented to the wind, the slope of the facet developed was much steeper than the angle at which the two facets of the other wind direction were cut (Fig. 2, illustrations 4 and 10); and this was so whether the oblique facets faced the north or the south wind.

With a rectangular stone, therefore, four facets are cut, corresponding to the side faces of the fragment. If the base of the stone is nearly square, then the theoretical development would be a low, symmetrical pyramid with four sloping sides meeting in a point; and further reduction, if the fragment was large and did not move, would be by progressive lowering of the height of the pyramid. In most natural cases the edge will not point directly into the eye of the wind but will occupy some position intermediate between that and the transverse. In such oblique positions two opposite side faces are more directly across the wind than the remaining pair; consequently, under favorable conditions, one pair tends to develop at the expense of the other. If, as we have already assumed in the previous case,

[9] Woodworth, *op. cit.*

the area of the base is originally large in proportion to the height, then the continued reduction of mass will produce a flattish, nearly horizontal plate. Such forms, usually with a thickness of $\frac{1}{8}$ inch to $\frac{1}{4}$ inch, are found among the Marlborough stones; but we have already noted (p. 205) that these are modified in the final stages by solution. Whether wind action alone is capable of reducing a fragment to a flat, horizontal surface is doubtful. Kirk Bryan[10] remarks that below 30° abrasion is slow and P. H. Kuenen[11] has expressed the opinion that there is a definite limit to the angle to which facets may be reduced. The writer's experience is that facets below an angle of 30° tend, by rounding of the interfacetal ridges, to become ill defined. Probably the ultimate result of wind action is a gently convex surface rather than a flat one.

THE INFLUENCE OF ORIGINAL SIZE

In the examples described above, all the original fragments either had, or were assumed to have had, a broad base in proportion to the height. In consequence of this, the opposite facets were reduced to a relatively low angle before meeting to form a ridge. The examples now to be considered differ in that the areas of the bases are relatively much smaller, so that complementary facets meet at a much earlier stage of their formation, producing more definite and sharper ridges between them. As a class they seem to be far more common than those of the previous group. It will be convenient to consider them in the same sequence as was adopted in the former discussion.

In the case of a rectangular stone so placed that it presents two faces at right angles to the directions of the two prevailing winds, two facets are cut facing in opposite directions. As the width of the base is not large, these two facets, by reducing the area of the upper surface, soon meet in a sharp ridge, to form the characteristic type of stone known as a *Firstkanter*. The further evolution of such a stone may, and usually does, differ considerably from that of a stone with a broad base and low-angle facets. Instead of continuing to rest on the base without disturbance while the angle of the facets is further reduced, pebbles with a narrow base commonly change their

[10] *Rept. Comm. Sedimentation, Nat. Res. Council* (1931), p. 35.

[11] *Leidsche geol. Medellingen*, Vol. III (1928), p. 31.

position. When a general critical size for the locality (a size governed largely by wind velocity) is reached, the pebbles roll over, presenting new surfaces on which fresh facets are cut.

The manner in which pebbles roll has, so far as the writer is aware, not yet been described from observation. Bather[12] conceived of the wind blowing away the surrounding sand, leaving the pebble on a pedestal, which was later cut through, causing the pebble to topple over into a new position. Kuenen, however, was unable to obtain any such result in his experiments, and no observation of such conditions in nature seems to have been recorded. J. W. Evans[13] was of opinion that the wind, by gradually removing the sand beneath and in front of the stone, caused it to fall, turning over toward the wind on its abraded surface. In this manner, he says, "a plane of abrasion [would] then be formed on the stone making an angle of about 60 degrees with the first." He believed that pebbles will thus advance *into* the wind. At precisely the time when the sand is most actively blown away, the wind will be at a maximum and will tend to prevent the pebble falling cleanly forward onto a facet. The writer was unable to find any specimens where this has undoubtedly occurred, and considers it far more probable that the pebble would merely sag toward the forming depression into an intermediate position.

As wind action continues on the stone in its new position, a second pair of facets is developed, the angle between them depending on the time and strength of the cutting, and their relation to the previous facets being governed by the amount of rotation of the pebble. Each movement, followed by cutting, will have the effect of reducing the size of the base on which the pebble rests and of making it more unstable. In this manner a "multi-faceted" form is developed, on which the facets are separated by parallel ridges after the fashion of a "zone" of faces on a crystal (Fig. 2, illustration 5). In the final stages, as each pause in a position of rest becomes shorter, a large number of small facets is characteristic and the stone approaches more and more a sub-cylindrical form (Fig. 2, illustration 8). Such pebbles are rolled over *with* the wind in the stronger gusts.

Rectangular blocks which are originally so placed that they pre-

[12] *Op. cit.*, p. 405.

[13] *Geol. Mag.*, Vol. VIII (1911), pp. 334–35.

sent an edge toward the wind will have the four side faces speedily beveled to produce a pyramidal form which will have much steeper facets than those on the pyramid-stones with a large base. These are typical *Pyramidalgeschiebe*. When such a stone topples or sags, a great variety of modifications becomes possible and the recognition of these later stages of pyramid-stones yet remains to be carried out. Fragments which occupy an oblique position may have two facets developed more strongly than the other two, and are very common in nature. Stones with two well-developed, opposite facets terminated by a smaller pair of facets have been described by many authors, and evidently owe their general occurrence to the greater frequency with which the originally oblique position occurs in nature.

In many instances where an original angle is found to be directed into the wind, it has been beveled and rounded to a smooth curve between the two flanking facets. This may be taken to indicate that the tendency of the sand blast is to smooth off such a forward angle, a conclusion which is in accord with the experiments of Kuenen. On the other hand, so many stones show the sharpest interfacetal angles pointing directly into the wind that the writer confesses that he has reached no conclusions based on observational data on this question.

THE INFLUENCE OF ORIGINAL SHAPE

Many writers, in discussing the cutting of facets and the development of wind-faceted stones, have paid no regard to the original shape of the pebble whatsoever, tacitly considering that only the wind direction governs the position of the facets developed. The extreme opposite view is taken by Heim and Kayser, the latter of whom has stated[14] that "the number and arrangement of [the] faces depends not upon the wind directions but solely upon the original shape of the pebbles." This was thoroughly tested in the laboratory by Kuenen,[15] who concluded that his experiments showed that "Heim's theory of the paramount influence of the shape of the basis before corrasion begins is confirmed."

The Marlborough occurrence is particularly favorable to a study of this problem. The material shed from the limestone outcrop not

[14] *Allgemeine Geologie, Enke*, Vol. I (1921), p. 311.

[15] *Op. cit.*, p. 26.

only exhibits stones in all stages of abrasion but also affords innumerable samples of fresh fragments. Owing to the bedded, "flaggy" nature of the limestone, most of these are in the form of flat slabs. Though rectangular shapes are common and have been taken as the basis of the foregoing discussion, many are triangular in form and others are polygonal.

Those which are triangular in plan show modifications under the influence of the sand blast which are in every way comparable with the forms already described. They fall readily into the same two classes: those with a relatively large base which, after a long period of wearing, exhibit low-angle facets (Fig. 2, illustration 10) and ultimately become almost flat plates (Fig. 2, illustrations 2 and 3); and those with a smaller base which develop into trigonal pyramids (Fig. 2, illustration 4) and later roll over to develop a large number of irregular facets.

Fragments which originally have a large number of side faces at first tend to have small facets cut on most of the original surfaces. Such forms, where the number of facets is determined mainly by the number of faces of the fragment, may be termed "vari-faceted," to distinguish them from the "multi-faceted" forms already described, where the number of cut surfaces is caused by later movement of the fragment. As some of the facets, by virtue of a more favorable position relative to the wind, may tend to become enlarged at the expense of others, the stone may undergo a reduction in the number of facets as wind action proceeds and tend to assume one of the shapes already described (tetra- or tri-pyramidal and *Firstkanter* forms are the most usual). In a region of light variable winds this might not be true.

From this it appears that, though wind action may modify the shape of a stone by reducing the number of facets and guiding it toward a more or less standard type, the original shape of the stone probably exercises an influence which is even more important in determining which of the more common types shall be developed. In this connection it may be remarked that pebbles in quite late stages of development frequently show some evidence of their original shape.

Rounded boulders may be expected to offer greater opportunity

for the facet positions to be governed by wind direction. A hurried examination of favorable material at the mouth of the Clarence River, 20 miles to the southwest, indicates that this is so, the number of major facets developed across the prevailing wind directions being nearly twice the number in the parallel position on ancient river-worn boulders.

THE INFLUENCE OF TILT OF THE BASE

In general, facets are developed in a normal manner upon pebbles the bases of which are not horizontal. Interfacetal angles are quite comparable with those cut on horizontally lying fragments, but the angle between any given facet and the base depends also upon the angle at which the base rests. A problem for laboratory study which arises here is the possibility of facets being cut beneath a stone when the forward edge is raised above the level of the surrounding sand.

CLASSIFICATION AND TERMINOLOGY

Kirk Bryan[16] discusses the various terms which have been applied to stones whose shapes have been modified by wind-driven sand. With many of his terms this paper has no concern, the only group studied being what he terms "wind-faceted stones (*Dreikanter*)." This group he subdivides into four types: "(1) irregular type, (2) pyramidal type (*Pyramidalgeschiebe*), (3) ridge-shaped type (*Firstkanter*), and (4) triquetous type (Brazil-nut type)." This classification is based wholly on the present appearance of specimens and takes little account of their previous history or probable subsequent evolution. The writer now proposes a somewhat more comprehensive scheme based on the factors described above:

a) As the size of the base of a rock fragment relative to its height primarily governs the subsequent form of the stone, determining whether it shall have low-angle facets developed upon it and ultimately be reduced to a low, flattish plate or whether it shall be characterized by steep facets meeting in sharp ridges and angles, a primary division may be made into two classes: *megabasids*, or broad stones with flattish facets, and *microbasids*, or stones with small bases and steep facets.

[16] *Rept. Comm. Sedimentation, Nat. Res. Council* (1931), pp. 30–31.

b) As the number of side faces on a fragment determines the number of facets originally cut and provides a shape which, persisting even to very late stages of cutting, may determine the form of the final product, prefixes such as di-, tri-, tetra-, etc., may be applied to the previous terms to indicate the number of facets present.

c) As the form of wind-cut stones is also influenced in a major degree by the stage to which abrasion has advanced, a secondary division may be made to indicate the influence of this factor. Stones show in their history a curious analogy with the dissection of a land surface. Commencing with a youthful stage, when the forms developed are only modifications of the original shape of the stone, they pass to an early mature form when the facets attain an angle of approximately 50°. When, by further reduction (at a slower rate) of facet-slope, opposite facets meet to obliterate the last remnants of the upper surface of the stone, the product may be termed fully mature; and later, when the facets are reduced to a very low angle, the stone may be deemed to have reached the stage of old age.[17]

Wind-faceted stones may thus be adequately described as: an early mature tri-megabasid (Fig. 2, illustration 10); a fully mature di- or possibly tri-microbasid (Fig. 2, illustration 6); or a youthful tri-microbasid (Fig. 2, illustration 4).

Though this scheme is apparently suited to the classification of forms with a wide base, certain difficulties arise when it is applied to fragments which possess a narrow base. First, in such a case opposite facets may meet before they attain the general angle which characterizes early maturity. The stone thus reaches late maturity first. This difficulty seems unavoidable. Second, when smaller pebbles turn over, permitting new facets to be cut, a complication is introduced which interrupts the normal sequence. If normal *Firstkanter* (mature di-microbasids) turn over, the multi-faceted type results; if *Pyramidalgeschiebe* (mature tri- or tetra-microbasids) turn, irregular forms are produced. When further work is carried out on the turning-over of these types, it may be possible to classify them more precisely; for the present they are left in this somewhat unsatisfactory state.

[17] Probably the ratio of the time intervals between "youth" and "maturity," and "maturity" and "old age," is also of the same order as in the topographical analogue.

CONCLUSION

Some of the features of faceted stones described above are possibly of only local significance. The dual action of the wind in this region, the restriction of the effects to limestone fragments only (the other local rocks disintegrate too rapidly to exhibit faceting), and the narrow width of the strand plain impose a set of conditions which may cause features to be developed which are quite out of harmony with those found elsewhere. It is possible that some of the principles enunciated have come to light only as a result of these special conditions, and may have no application in regions where factors of wind velocity and direction and the constitution of both pebbles and abrading material bear no comparison with those of Marlborough. For the investigation of these points the writer can only depend upon the kind offices of colleagues in other favorable localities.

THE EVIDENCES OF WIND ABRASION

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ABSTRACT

In a recent summary of ventifact localities in the United States, the omission of certain regions where wind abrasion is widespread and obvious indicates that the criteria of this process are often overlooked. Conspicuous examples of wind carving are to be found in the Dells region of Wisconsin and the upper Arkansas River valley and South Park in the Rocky Mountains of Colorado. Little wind work now occurs in these regions; most of it probably took place under periglacial conditions in the Pleistocene ice age.

In a recent paper Wentworth and Dickey[1] enumerated the ventifact localities in the United States disclosed by a rather wide inquiry. The omission from their list of certain localities where wind abrasion is conspicuous and widespread leads the present writer to believe that criteria of this process are not always recognized by the field geologist. Some of the localities that have been overlooked are briefly described herein.

Three regions where abundant evidences of wind abrasion are to be found are: (1) the sandy plain of the glacial Lake Wisconsin, extending for many miles north and west of the Dells of the Wisconsin River in the state of that name; (2) the valley of the Arkansas River between Leadville and Salida in the Rocky Mountains of central Colorado; and (3) the broad semiarid intermontane basin known as South Park, Colorado.

The former Lake Wisconsin was a shallow ice-barrier lake formed in the upper Wisconsin River valley at the Cary or third substage of the Wisconsin glaciation. Chief among the evidences for its existence are erratic crystalline boulders supposedly rafted by ice to their present positions within the limits of the lake basin and up to its highest level of 960 feet. Some of these crystallines up to 2 feet in length, beautifully and unmistakably grooved by wind-blown sand, are found north of the Wisconsin River in the west half of Section 3, T. 13 N., R. 6 E., Dells topographic quadrangle map. These boul-

[1] Chester K. Wentworth and R. I. Dickey, "Ventifact Localities in the United States," *Jour. Geol.*, Vol. XLIII (1935), pp. 97–104.

ders lie on the southeast side of a hill of sandstone against which their icebergs doubtless stranded and upon melting deposited them. The wind grooving, although sharp and unmistakable, is not in the process of formation at the present time, and thus these stones are to be classed with Wentworth's "fossil ventifacts."[2] Ventifacts are numerous a few miles northward at the northwest base of the hill known as Elephant's Back in Section 21, T. 14 N., R. 6 E., also within the sandy Lake Wisconsin plain. Here many pebbles of crystalline rocks and limonitic crusts derived from the weathering of the Potsdam sandstone bedrock are grooved and brightly polished by the action of wind-blown sand. On one sandy knoll deflation has so removed the finer material that these wind-carved pebbles have accumulated at the surface as a true "desert pavement."

Although some movement of loose sand is now carried on in this region by the wind, such movement was surely very much greater in early postglacial time, when wind velocities were high and before vegetation had covered the sandy soil of the Lake Wisconsin plain.

Evidences of wind abrasion in the Arkansas River valley are best developed between Leadville and Salida. About 25 miles below (south of) Leadville the river emerges from a narrow crystalline-rock gorge into an intermontane basin several miles broad, eroded in Pliocene clay and sand. The minor topographic forms within this basin belong mainly to two types: (1) glacial moraines of the Wisconsin and at least one earlier glacial stage, and (2) broad terraces ranging in age from late Pliocene to postglacial. Near the north end of the basin these terraces are strewn with large crystalline boulders up to 15 feet in diameter, far larger than any now handled by the Arkansas River and evidently brought to their present positions when the river was greatly augmented in size. Most of these great boulders are to be found on the terraces belonging to the glacial stages, and thus mark the river's increase in carrying power due to glacial melt-water. These boulders show abundant evidences of wind abrasion, mainly in the form of grooves and flutings which indicate wind movement from the northwest. These markings do not cover the boulders at present, but occur in certain patches or crusts where recent weathering has failed to remove them. Thus they point

[2] *Ibid.*, p. 102.

to a climatic period when winds were stronger and sand particles less fixed by vegetation than at present. One thinks at once of the severe periglacial conditions that must have existed in this valley when the last glaciers occupied the slopes of the Sawatch Range near by.

The youngest and lowest river terrace, developed in postglacial time, is less extensive than terraces belonging to the Wisconsin and pre-Wisconsin stages. Because the wind-grooved crusts are so numerous on the boulders of the glacial terraces, particularly those older than the Wisconsin, the writer at first regarded the wind-cutting as a criterion of their glacial age. However, careful search disclosed feeble wind carving about the lower parts of boulders on the postglacial terrace and abundant grooving on stones of the Wisconsin terrace and even parts of the Wisconsin moraine. Therefore wind grooving cannot be used in fixing the age of physiographic surfaces on which the ventifacts occur, though it is much more strongly developed on relatively old surfaces than on those of Wisconsin age or later.

Although grooves are the characteristic form of wind-cutting here, a few of the smaller boulders show pits and "collar-stud" points on the northwest side, against which the sand had been driven (Fig. 1).

In South Park, Colorado, wind abrasion is so widely evidenced that conditions must have been extremely favorable for its development. South Park is a broad depression of which the surface features consist of: (1) extensive terraces or pediments, in various stages of dissection but with many flat areas of great extent, and (2) certain minor hills and ridges up to a few hundred feet in height. Due to the semiarid climate, trees are practically confined to the higher ridges and a few favorable valleys, and the greater part of the Park area is therefore treeless. The cover of short grass, sage, and certain other small shrubs is insufficient even now to prevent the movement of loose sand by the wind. Certainly the vegetative cover was even thinner during the several glacial stages, for when glaciers occupied the lofty Mosquito Range and other mountains on the west and north, periglacial conditions must have prevailed throughout the Park. Only the lowest and least extensive of the terraces was formed in the Wisconsin stage of glaciation; all the higher terraces must

therefore have suffered more than one period of severe periglacial conditions.

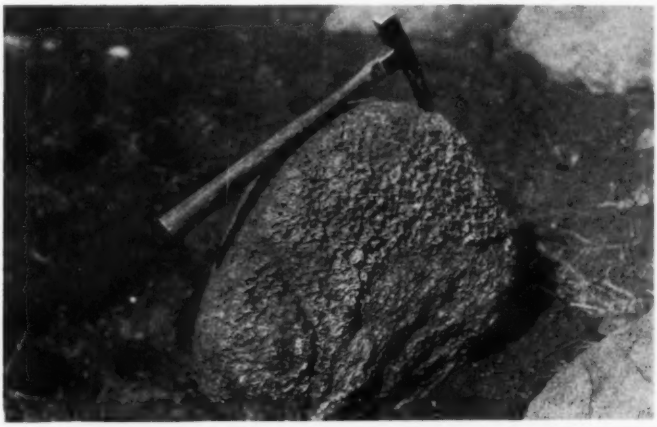

FIG. 1.—Surface of pits and "collar-studs" developed by windblown sand. Boulder lies on Wisconsin outwash terrace in Sec. 19, T. 13 S., R. 78 W., upper Arkansas River Valley.

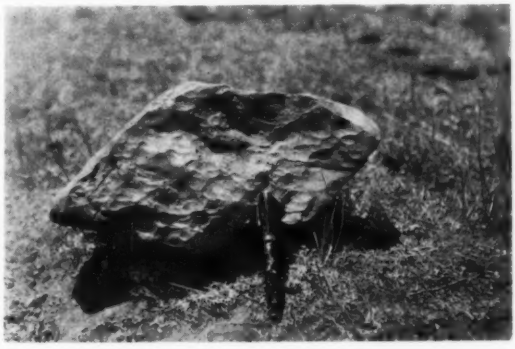

FIG. 2.—Grooved surface of boulder shaped by wind. Stone rests on surface of terrace probably of very early glacial age, in Sec. 12, T. 13 S., R. 76 W., South Park.

In such an area one would expect to find evidences of wind carving —and these are abundant. The surface boulders and gravel of the terraces show grooving and pitting (Fig. 2), and many of the smaller

stones have had their edges sharpened in such a manner as to suggest einkanter and dreikanter, though true forms of these types are rare. Stones partly embedded in the soil generally are grooved only on their exposed portions, and especially on their northwest sides. Ledges near the base of the volcanic hills at the south end of the Park also exhibit beautiful grooves and flutings. The average trend of such grooves in a large number of determinations on rock ledges and undisturbed surface boulders was found to be N. 60° W., indicating wind movement from that direction.

FIG. 3.—Wind-grooved crust on granite cliff now in process of exfoliation. Windstream moved from left to right, *from* N. 60° W. Sec. 4, T. 9 S., R. 75 W., South Park.

Although it seems clear in theory that these wind carvings belong mainly to a past period of strong winds and sparse vegetation, some of the features themselves give positive evidence of their ancient origin. Figure 3 shows a set of grooves preserved upon the side of a granite cliff or ledge in Section 4, T. 9 S., R. 75 W. The grooves are part of a thick crust some two inches thick that elsewhere on the ledge has been removed by exfoliation. Thus the grooved crust is not now in the process of formation, but rather of destruction. Furthermore, the cliff forms the edge of a granitic upland rising 75 feet above a grassy lowland on the west. The present conditions effectively prevent abrasion by westerly winds; therefore the grooves must have been cut before the lowland on the west was etched to its present depth, i.e., before the Wisconsin glacial stage. Surface stones on the

higher and older terraces have suffered much stronger wind abrasion than those on the Wisconsin outwash terrace. This fact also indicates that much of the carving is pre-Wisconsin.

CONCLUSION

That wind abrasion has not hitherto been reported from such regions as the upper Arkansas River valley and South Park, Colorado, indicates that the criteria of this process are often overlooked. Rock surfaces shaped by wind abrasion are distinctive and unmistakable. Angular edges parallel to the wind stream are left even sharper as the faces on either side are worn concave by the sand blast. Flat surfaces against which the sand is squarely driven become eroded differentially and develop pits, projecting pinnacles, and "collar-studs" (Fig. 1). Flat surfaces along which the wind sweeps obliquely are generally grooved. The grooves are out indiscriminately across minerals of different hardness, and consist of shallow, rounded, nearly parallel channels separated by relatively sharp divides or ridges (Fig. 2). It is the writer's observation that such forms as einkanter and dreikanter, although highly distinctive, are far less characteristic of wind abrasion in the United States than the other forms mentioned above. Failure to find einkanter and dreikanter may have led many to overlook other obvious criteria of wind abrasion.

It is also likely that wind abrasion has remained unrecognized due to lack of understanding of past climatic conditions. Where today a heavy vegetative cover prevents any significant movement of surface sand by the wind, evidences of active wind-erosion under a former different climatic régime are likely to be overlooked. It is the writer's opinion that most existing wind-abraded surfaces in the three regions herein discussed were formed under periglacial conditions of strong winds and greatly reduced vegetative cover.

REVIEWS

Introduction to Geology. By E. B. BRANSON and W. A. TARR. New York: McGraw-Hill Book Co., 1935. Pp. 470; figs 456. $3.75.

No teacher of geology need say that his colleagues have not provided him generously with (fairly) satisfactory texts. To the writer, the only perfect text is LeConte's *Elements*, seen through the years in the rosy light of early geological enthusiasm.

There is no question of the scientific competence of the authors of any of the geology texts in common use. The value of the book is almost wholly a matter of choice of material, clearness and logical order in presentation, and skill in choice of photographs and in construction of drawings.

This text covers both physical geology (312 pages) and historical geology (151 pages), without possibility or desirability of a two-volume edition. It seems designed for a one-semester intensive course, in which a brief survey of historical geology is included. Where but one course in geology is taken by the student, such an inclusion seems eminently desirable. The book could, however, be used for a year course. No dogmatic statement here will hold, for instructors in geology add (or subtract) from any text according to individual preference, and especially according to the region in which they are working.

The historical part, entirely by Branson, divides as follows: pre-Cambrian, 10 pages; Paleozoic, 58 pages; Mesozoic, 45 pages; Cenozoic, 18 pages; Pleistocene, 13 pages, of which 5 deal with man. This may be in proportion to the lengths of the periods, but hardly does justice to the last two, during which the major features of American scenery were shaped. The chief emphasis is on life, exceedingly well done. Occasional side lights on collecting and collectors add to the interest. The great physical revolutions, Appalachian and Laramide, might well have received fuller treatment.

The part dealing with physical geology starts with minerals, igneous rocks, and vulcanism; goes on to weathering (of igneous rock), to surface agencies, rivers, ground water, ocean, ice, and wind; and ends with diastrophism, earthquakes, and the earth as a whole. Sedimentary and metamorphic rocks are intercalated between ocean and ice. Here again the instructor is free to follow his own order. The development of the

successive chapters is full, orderly, interesting. The authors say in the preface that "perhaps the greatest debt of the authors is to their students, who have shown them the way toward improvement in the selection and presentation of the materials." This is borne out in many details of the text. Worth mentioning is the frequent, often quite incidental but not less effective, introduction of practical applications of geology to economics and engineering.

A careful reading of selected sections shows occasional infelicities, even inaccuracies, of statement, which should have been caught in a careful word-by-word examination of the completed manuscript. Speaking for users of textbooks, both instructors and students, it would be an excellent thing if all writers of textbooks could see desk copies of their texts which have been annotated by instructors using them, and could use the suggestions in preparation of later editions.

Many of the new photographs are excellent. Some have been "retouched" to an extent that makes it difficult to tell whether one is looking at a photograph or a drawing. A number of the diagrams are of special value. Shaded drawings are not so satisfactory as line drawings. Devising illustrations is an art in itself. Some, like Davis' of mountain glaciation, Johnson's of the shore cycle, and now of Cenozoic evolution of the Appalachian region, are so good that the student is entitled to them in any text if it is possible to get them. No substitute will "do just as well."

Two suggestions apply no more to this text than to others, but they are worth making here. The first is to abandon the irritating method of referring to illustrations by an independent series of numbers; they should be referred to as, for example, "Fig. 21*A*," where the number is the page number, and the letter is used only where there is more than one figure on the page. This would make it much easier to find the illustration. The other suggestion is to put more of the description of the illustration immediately with it and less in the text. Students using a textbook should study the illustrations as carefully as the text; putting illustration and text together will aid this. The present text carries the principle farther than some, and could carry it much farther with advantage to the student.

All in all, Branson and Tarr have produced a good text. It will be interesting to watch the struggle for existence among textbooks in geology which come from Yale, Princeton, Chicago, Minnesota, Southern California, and now Missouri. On the "multiple working hypothesis" basis the more produced the better the outlook for the student of geology.

LEWIS G. WESTGATE

Cosmogonies of Our Fathers. By KATHERINE BROWNELL COLLIER. New York: Columbia University Press, 1934. Pp. 500. $5.00.

Here is a book that will interest many geologists. The authoress traces, in chronological sequence, the manner in which the scientific discoveries of the seventeenth century were reconciled with scriptural accounts. Among these reconciliations were many that figure prominently in the history of geology. Theories to account for the origin of the earth and its present configuration are given in sufficient detail to make the volume an important reference for geologic historians.

The book begins with a discussion of the conflict of systems of thought, and at once takes up the ideas of Fludd, Descartes, Kircher, Burnet, and Woodward, to mention but a few. Each of the writers included is treated fully, and complete references are given to edition and page. Following the chronological discussion of writers is a section devoted to such topics as the heavens; the classification of matter; earthquakes and eruptions; mountains; metals and minerals; spontaneous generation; etc. A comprehensive Bibliography and a detailed Index close the volume.

Although the book is quite broad in its general conception, there are many topics pertinent to geology in it. Certainly it is a volume that should grace the reference shelves of geologists interested in the historical development of their science.

W. C. KRUMBEIN

Lehrbuch der angewandten Geophysik. By HANS HAALCK. Berlin: Gebrüder Borntraeger, 1934. Pp. vii+376; figs. 142; tables VI. Rm. 24.

This volume of geophysical prospecting methods will appeal especially to the student of practical geophysics. The treatment is well rounded, involving as it does both the underlying theory and the equipment needed for practical application. The apparatus used is illustrated with many halftones and diagrams.

Several parts comprise the book. The first section discusses gravimetric methods; then follow magnetic and electrical methods. Seismic methods of prospecting also receive detailed treatment, and the volume closes with a section on radioactive measurements. In each section numerous practical examples are given, including in some cases the actual field data, supplemented by maps and diagrams.

The text is well written throughout and the figures and halftones are clear. Several isogonal maps and like data are included at the end of the volume.

W. C. KRUMBEIN

Paläogeographie. By FRITZ KERNER-MARILAUN. Berlin: Gebrüder Borntraeger, 1934. Pp. viii+410; figs. 27. Rm. 32.

Instead of a descriptive discussion of paleogeography, the author has written a volume which treats the subject analytically, and investigates the basis on which paleogeographic research is founded. Each of the devices and methods used by paleogeographers is discussed and critically examined. The treatment is excellent, and the book fills a need for such a presentation of the fundamentals of the science.

The first section of the work is devoted to sedimentary petrography, from the viewpoint of using sedimentary data in reconstructing the past. The criteria used and the conclusions based on them are given, as are the sources of error that may enter the work. Numerous examples are touched upon here, including the significance of pyritiferous shales, salt beds, and the like. This section is followed by one on correlation, involving the theories supporting it, and the variables that may enter the problem. Theories of mountain building, in so far as they touch upon the general subject, are well presented in a following section, and the volume closes with a section on paleocartography itself. Here are included maps of terrains rising or sinking relative to sea-level, as well as conventional diagrams and maps. As may be expected, the illustrations and examples are drawn largely from European sources.

The books deserves the consideration of geologists in both the physical and biological aspects of the science. It will prove a valuable reference for those engaged in summarizing historical geologic data on maps.

W. C. KRUMBEIN

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BORE-HOLE INVESTIGATIONS IN YELLOWSTONE PARK

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ABSTRACT

In connection with the study of thermal phenomena of Yellowstone Park carried out by the Geophysical Laboratory for several years, two bore holes were drilled in geyser basins. During the progress of this work, temperatures and pressures were measured and observations were made on underground structures shown by the cores. Subsequently the cores were studied in the laboratory by chemical and petrologic methods to ascertain the mineral transformations effected by hydrothermal circulation, and their significance.

Structural features, such as porosity and fractures, together with thermal conditions found, help to explain the underground circulation. The data enable us to picture the manner in which meteoric waters, falling upon adjacent plateau areas, penetrate to deep levels, are met by magmatic exhalations at high temperature, are diverted upward, and appear at the surface as hot springs and geysers. Light is thrown upon the mechanism of geyser action itself, which is found to be a rather more complicated process than has been supposed.

Metasomatic alteration of the rocks has been important. A considerable variety of new minerals has been formed, and the composition has been much changed. The most notable effect found in the cores has been the addition of silica and the molar replacement of soda and lime of the feldspars by potash. The results of this study supplement and confirm those obtained by E. T. Allen in his study of the composition of the thermal springs, carried out as part of the same project. From this assembly of concordant data we are enabled to trace out with considerable confidence a series of reactions that apply not only to the rocks rendered accessible by the drill holes but to those at greater depth, and to deduce the manner in which the original magmatic emanations, dissolved in meteoric waters, have modified the composition of the rocks through which they have passed, and in turn have been modified.

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INTRODUCTION

For a number of years several members of the Geophysical Laboratory were engaged in a comprehensive study of the geysers and hot springs of Yellowstone Park. A description of results obtained in many fields of investigation has lately appeared.[1] Short abstracts of special features have also been published from time to time.[2]

[1] E. T. Allen and Arthur L. Day, *Hot Springs of the Yellowstone National Park*. Microscopic examinations by H. E. Merwin. Publication No. 466, Carnegie Institution of Washington, 1935.

[2] E. T. Allen, "Neglected Factors in the Development of Thermal Springs," *Proc. Nat. Acad. Sci.*, Vol. XX (1934), pp. 345–49; "The Agency of Algae in the Deposition

As part of this investigation the temperatures of hot pools and geysers were measured to such depths as were accessible from the surface, but it was usually not possible to explore these pools to greater depths than 40 or 50 feet, because of constrictions of the channel or deviations from the vertical. In order to obtain a better understanding of underground structure, temperature, and circulation in geyser basins, features on which surface manifestations are obviously dependent, the drilling of bore holes to a depth of several hundred feet at selected sites was decided upon.

Arrangements for the boring operations were made with a driller-man, Vance Butler, who had had much experience in drilling steam wells in California and elsewhere. Sites were chosen and methods of procedure were formulated by Day and Allen. The direct supervision of operations in the field, and the recording of information, were placed in charge of the present writer. The subsequent petrographic and chemical studies of the cores were also carried out by the writer. These studies were found to be of assistance in interpreting processes that gave to the surface waters some of their important characteristics, and such results have been incorporated in the comprehensive volume mentioned. Certain of them have also been published in brief abstract.[3] Other results, especially those of more strictly geological and mineralogical import, have not been published. It seems desirable to bring together in a single article various results of this kind in their interrelations.

In our investigations, surface and underground features of the hydrothermal circulation of the Park have been found to be so closely interrelated that discussion of one subject necessarily involves the other, if its significance is to be made clear. Therefore, this article is to some extent the presentation of another aspect of the phenomena described in the more comprehensive volume by Allen,

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of Travertine and Silica from Thermal Waters," *Am. Jour. Sci.*, Vol. XXVIII (1934), pp. 373–89; "Geyser Basins and Igneous Emanations," *Econ. Geol.*, Vol. XXX (1935), pp. 1–13; E. T. Allen and Arthur L. Day, "Hot Springs of the Yellowstone National Park," *Proc. Fifth Pacif. Sci. Congress*, Vol. III (1934), pp. 2275–83.

[3] C. N. Fenner, "Hydrothermal Metamorphism in Geyser-Basins of Yellowstone Park, as Shown by Deep Drilling," *Trans. Amer. Geophys. Union, 15th Annual Meeting* (1934), pp. 240–43.

Day, and Merwin, but in most respects from a quite different point of view, and with emphasis on other features.

Previous to the drilling of the holes the present writer had assisted Day and Allen in some of their investigations in the Park, and had discussed with them the significance of phenomena. Discussions were continued during the later studies of drill-hole material. With Allen, especially, were the results of his chemical investigations of waters and gases, and their relation to processes of metamorphism in the rocks, fully considered. Certain petrographic features were discussed with Merwin, particularly the identification of clays that he had studied in surface samples collected by Allen, and their relation to the clays in Norris drill cores. From all these investigators the writer received most useful information. The cordial co-operation and assistance given by the officials and ranger staff of Yellowstone Park greatly facilitated field operations and were much appreciated.

Two holes were drilled, the first in the fall of 1929 in the Upper Geyser Basin, or Old Faithful Basin as it is commonly called, and the second in the fall of 1930 in Norris Basin.

THE BORE HOLE IN THE UPPER GEYSER BASIN; GEOLOGICAL SETTING AND NARRATIVE OF DRILLING

In the Upper Geyser Basin a site was selected by Day and Allen to meet several requirements. Some apprehension had been expressed that the drilling of a hole would interfere with the underground circulation and disturb geyser action, and though this fear was believed to be unwarranted, it was decided to operate outside of the main geyser areas, but still within the undoubted region of hydrothermal activity. Also, it was well to choose a site somewhat remote from the chief points of tourist interest.

The site selected was several hundred yards westerly from Old Faithful Geyser (a little south of the Three Sisters group), in a wide, nearly level plain, whose surface is covered with fragmental sinter (see Fig. 1). Numerous small hot pools were in close proximity, and some of these showed at times a little spouting or periodical overflow.

Study of exposures in the neighborhood indicated that all the surrounding area is underlain by a series of roughly stratified gravels and sands. Figure 2 shows a typical exposure at a gravel pit a few

FIG. 1.—Drilling rig in Upper Basin. View looking east of north

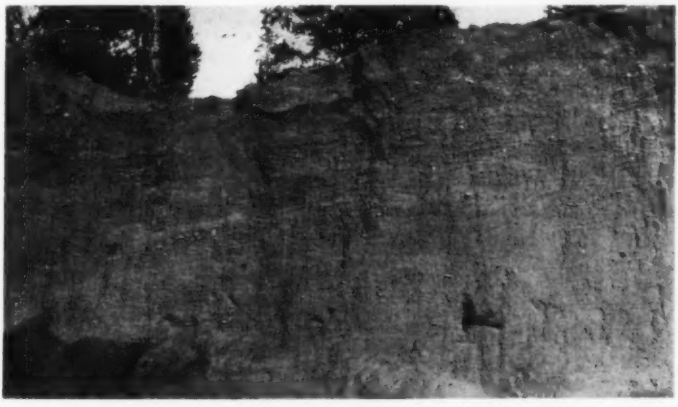

FIG. 2.—Irregularly stratified gravel and sand in an excavation not far from the drill site in Upper Basin.

hundred yards away in a northerly direction. This excavation had a maximum depth of about 25 feet. The material was an unconsolidated medium gravel with some larger pebbles and a few cobbles. It consisted almost entirely of obsidian and rhyolite, well stratified, but the strata were of wavy form and in places showed distinct cross-bedding. Pebbles and grains were rounded to subangular. Many had glassy opal adhering to them or cementing them in a way which in-

FIG. 3.—Terrace bounding east side of Lower Geyser Basin, a little north of Rabbit Creek, looking easterly.

dicated deposition *in situ*. Warm water was said to have been met in the deepest part of this excavation.

At other places in the vicinity and over a much wider area natural exposures of similar gravels occur, some consolidated to hard rock by deposition of opaline silica. There is good reason for supposing that sedimentary deposits of like character extend over nearly the whole area of Upper and Lower Geyser basins. In many places in both basins these gravel deposits form broad terraces (Fig. 3), which could hardly have been developed otherwise than through the agency of large bodies of water. A few low ridges of rhyolite protruding through the gravels indicate considerable relief of the buried rock floor.

From these observed relations we expected that the upper part of the drill hole would penetrate gravel beds, but the thickness of the beds actually found was rather astonishing.

Drilling was begun about the middle of September, 1929. An annular drill-bit of "hastellite" alloy was used, which brought out a core of about $1\frac{3}{4}$ inches diameter.

For a few feet downward from the surface the drill passed through siliceous sinter, of the sort that the geysers and hot springs now deposit around their orifices and in their overflows. At a depth of about 7 feet pebbles of rhyolite began to appear, and the proportion gradually increased. At 40 feet the core consisted essentially of gravel cemented with opaline silica and secondary quartz, and from here downward the gravels were thoroughly indurated. In this upper portion of the drill hole the water was warm but was not under pressure. At a depth of 62 feet steam was first encountered. As the hole advanced, the increasing pressures were measured with a Bourdon steam gauge, and temperatures with a maximum thermometer, lowered to the bottom of the hole.

In the upper portions of the hole, where the cementing material was chiefly opaline silica, the silica filled the interstices of a mass of gravel. The deeper cores were carefully examined to see whether they contained interstratified layers of sinter or fragments of sinter such as occur abundantly at the surface wherever deposits are forming at the present time. Their presence would have indicated that the gravels had probably accumulated subaerially and slowly enough to allow sinter to be deposited on temporary surfaces. No such fragments could be found. This might be supposed to mean that the induration of the mass of gravel occurred entirely after its accumulation, but it is questionable whether this inference is justifiable. The gravels probably accumulated under a large body of water. The effect of dilution when thermal springs emerge into a body of water is not well understood. We know that sinter is sometimes deposited under such conditions, but in other instances dilution seems to prevent deposition. It is possible, also, that in the case with which we are dealing opaline silica was deposited in and on surface layers of gravel and later became transformed to quartz.

The pebbles and grains of these gravel beds were originally of

such composition as would be derived from the mechanical disintegration of rhyolite. Even in their present generally altered condition they include some unaltered lithoidal rhyolite, obsidian and perlitic obsidian, spherulites, and phenocrysts. A maximum diameter of pebbles is ordinarily less than one-half inch, though some are larger than one inch. From these dimensions they grade downward to a fine gravel, sand, or silt. Most of the pebbles are well rounded, but somewhat angular pieces occur frequently.

In the deeper parts of the drill hole the gravels have undergone thorough mineralogical changes. In places the metamorphism has progressed so far that the original characteristics of the beds have been obscured, and it would be difficult to recognize the nature of the resulting rock if it were not for clues afforded by transitional stages. These changes will be described in detail in the section devoted to petrography.

A faint but distinct odor of H_2S was perceptible during the drilling.

Altered sediments continued downward to a depth of 220 feet before rock in place was met. This great thickness of sedimentary beds carries certain implications regarding the history of the geyser basins. From the topographic map of the United States Geological Survey it is seen that the surface elevation at the drill site is about 7,320 feet. The elevation of the rock surface beneath the sediments is therefore about 7,100 feet. The Upper and Lower Geyser Basins are surrounded by high hills of rhyolite (or rather, the basins are depressions in a rhyolite plateau), and the only outlet for the drainage of the basins is through the narrow, rocky valley of Firehole River. The Firehole enters this cut in the rock rim about 8 miles to the north of the drill site. The elevation at this outlet is about 7,150 feet. That is, the rock floor of the Upper Basin at the drill site is no higher than the rock floor of Firehole River 8 miles to the north.

There is no reason to suppose that the drill hole encountered the rock surface at its lowest point in the Upper Basin; in fact, it seems not unlikely that it may be much deeper in places.

We recognize that these relations are not such as are produced by normal stream erosion, and we are inclined to attribute them to overdeepening by glacial scour. The gravel beds have the characteristics that might be expected of sediments deposited in lakes ponded by glacial ice.

The formation of the basins themselves, and the length of the period during which hydrothermal activity has persisted, present an interesting problem. In regions of less abnormal phenomena such depressions would suggest the sinking of faulted blocks, but in Yellowstone Park the influence of thermal activity upon rock decay and erosion must be kept in mind. Many instances have been noted by the writer of what seemed to be more than chance association of hot springs and fumaroles with topographic depressions—basins and their small tributary valleys as well as main valleys—where no direct evidence of faulting was perceptible. Possibly a combination of several factors has brought about these results.

After the sediments were drilled through at a depth of 220 feet the rock encountered was originally of dacitic composition as far as the hole penetrated, 406 feet. Some of it had unquestionably been obsidian, or pitchstone, and a great part of it may have been of this character, but some may have been lithoidal rhyolite. Certain "islands" were passed through, which had been very little altered, but most of the rock was much changed.

Mention has been made of small natural pools of hot water in the neighborhood of the drill site. It was of interest to see whether the underground water circulation was so interrelated that drilling operations would have an effect upon them. When the drill hole was 300 feet deep, steam was allowed to blow off with the valve wide open for an hour. The volume of condensing steam at the surface was so great as to give rise to a considerable flow of water, but the natural pools 50–75 feet distant to north and south were not affected by this relief of underground pressure.

At a depth of 375 feet, however, a minor mishap occurred in drilling, by which several lengths of drill tube dropped to the bottom of the hole. In recovering them, it was found necessary to let steam escape without restraint for many hours. Apparently as a result of this a small hot pool about 50 feet distant was disturbed. Formerly the water level had been observed to rise periodically to the rim and sink slowly back to a depth of 22 inches. Simultaneously with the trouble in the drill hole, the water began to rise rapidly at intervals and overflow in a rather large stream, and then sink back. Whereas it had been clear, it now became slightly turbid, and bubbles of gas were given off. Another pool, 75 feet to the east, was not affected,

but in three very small pools to the north the water level rose higher than it had stood previously, and then subsided. Other pools at a greater distance were undisturbed.

When a depth of 406 feet was attained in drilling we believed we had obtained most of the information that was sought. Work was discontinued, and the orifice was sealed with concrete.

The percentage of core recovery during the drilling would not be considered altogether satisfactory in drilling oil wells, where the identification of each stratum is desirable, but it was probably sufficient to give us the required knowledge as to structural features and general character of rock and the mineralogic changes it had undergone. The amount of core recovered varied greatly. In places where the recovery was small this seemed to be due to the porosity and incoherent character of the altered rock. At intervals the rock became so hard that the "hastellite" bit was unable to make progress through it, and a chopping bit was used. In such cases, examination of the fragments washed out indicated that hard veins had been met or that the whole rock had become so impregnated with secondary minerals as to form a hard, dense mass.

TEMPERATURES AND PRESSURES

We believe that the successive temperatures measured at this boring represent a fairly close approximation to what would have been found if it had been possible to measure temperatures without drilling a hole. Though some disturbance was caused by the circulation of feed-water and by other incidents of drilling, these were probably rather temporary in their effect. As regards the measured pressures, however, a different interpretation is indicated. If it had been possible to measure the steam pressure at the same point as that at which the temperature was taken, it should have corresponded to the pressure of saturated water vapor at the given temperature, but it was always less, and, at times, much less. The highest reading was 57 pounds per square inch, to which should be added about 15 pounds for atmospheric pressure not taken account of by the gauge; whereas the highest temperature of 180° should give double this pressure. We found, moreover, that pressure readings taken before and after letting steam blow off usually differed considerably.

Irregularities of this kind, and the low value of all measurements, are presumably due principally to the fact that pressures were measured, not at the bottom of the hole, but at the surface, and the steam pressures were those at the top of a column of hot water that entered the hole at various levels when the valves were closed.

In a final section of this paper a discussion will be given of the system of circulation in the geyser basins, but it may be well to outline here our conception of some of its features.

We believe that the primary source of heat is a large igneous intrusive, which is cooling and giving off vapors. Meteoric water falling upon the adjacent plateau region penetrates to very considerable depths and moves laterally into the geyser basins, there absorbs heat from the uprising vapors and condenses them, is thus deflected upward, and issues again in geysers and hot springs. The quantity of meteoric water in this circulation decreases with depth, and at some depth the integrated result of heat absorption is just sufficient to take care of the magmatic heat. This depth is the level at which condensation occurs. The heat received is carried to the surface and is dissipated as steam or removed by stream discharge.

In the hole in Upper Geyser Basin the temperature gradient was found to be fairly regular (Fig. 4), and it might appear reasonable to extrapolate the graph to greater depths, and thus calculate the probable depth to which it would be necessary to go to obtain a temperature of, say, 1,000°. Extrapolation in this manner shows that not until a depth of 5,000 feet is attained would the temperature reach 1,056°. This calculation, however, probably has no significance. We might, with equal justice, assume that a batholith at high temperature lies only one or two thousand feet below the bottom of the hole, and that the upward transfer of heat from this body by gas convection and by rock conduction encounters at a certain level the downward penetrating and laterally moving ground water, and that here a great decrease of temperature occurs within a short interval. This assumption corresponds more nearly to conditions as we picture them, though the depth of the plutonic body is unknown.

The temperature gradient within the zone penetrated by the drill holes is the resultant of two independent factors: the amount of heat given off by the magmatic body and the volume of meteoric

water by which it is carried away. At some unknown depth penetration of meteoric water ceases, and below this the gradient is probably very different.

We give here a tabulation of the temperatures and pressures found at various depths in the drill hole. For reasons explained, the pressures are not regarded as indicating steam pressures at the depths given, and therefore they are not highly significant, but the variations are of some interest as indicative of the complexity of the system of channels tapped.

The first evidence of steam pressure was at a depth of 62 feet. The steam probably came in through a crevice, as it was encountered suddenly.

At 95 feet a pressure of 13 pounds per square inch was measured. (All pressures are the amounts by which atmospheric pressure was exceeded.) The temperature here was $125\frac{1}{2}°$.

At 104 feet the pressure was 12 pounds, temperature 129°. After allowing steam to blow off for a few minutes, pressure was $27\frac{1}{2}$ pounds. After blowing off a second time and again after a third time the pressure was 27 pounds. Final temperature $127\frac{1}{2}°$.

At 139 feet pressure = 25 pounds, temperature, 132°. After blowing off, pressure = 30 pounds; after a second blow-off, 29 pounds.

At 148 feet pressure = $32\frac{1}{2}$ pounds, temperature = 145°. After blowing off, pressure = $37\frac{1}{2}$ pounds.

At 156 feet first pressure = 37 pounds, second pressure = $37\frac{1}{2}$ pounds; temperature = 147°.

At 167 feet first pressure = 37 pounds, second pressure = $37\frac{1}{2}$ pounds; temperature = 146°.

With hole at 195 feet the pressure was $27\frac{1}{2}$ pounds. The thermometer gave a reading of $146\frac{1}{2}°$, but it was found that it had been hung up by an obstruction at 121 feet. A device had to be prepared for assuring that the bottom of the hole would be reached.

At 205 feet temperature = 155°, pressure = 28 pounds. The valves were opened wide and steam allowed to escape for 20 minutes. Pressure then 33 pounds. The drop in pressure from 37 pounds at 167 feet to 28 pounds at 205 feet, and similar irregularities elsewhere, should be ascribed to the complicated system of fractures cutting the rock.

When the bottom of the hole was at 220 feet another trial was made of letting the thermometer down on a wire. It was halted at 121 feet as before, and we noticed that there was a plainly perceptible jerking of the wire, as if a forcible jet of steam or water were striking the thermometer. The temperature registered 141°. The steam pressure before blowing off was 32½ pounds, after blowing off, 37 pounds.

With bottom of hole at 226 feet the temperature at bottom was 159½°.

At 242 feet a pressure reading taken before blowing off steam gave 31 pounds, after blowing off for several minutes, 40 pounds. It then rose to 42½ pounds, and after 7 minutes it had dropped to 37 pounds.

Similar variations of pressure continued in the lower parts of the hole, but they will not be listed in detail.

At 264 feet temperature = 164½°.

At 292 feet temperature = 171°, pressure = 42½ pounds.

At 313 feet temperature = 175°, pressure = 42½ pounds.

At 343 feet temperature = 173½°, pressure = 43 pounds.

At 379 feet temperature = 172°, pressure = 35 pounds.

At 406 feet temperature = 180°, pressure = 49 pounds, before blowing off, 57 pounds after.

In the accompanying diagram (Fig. 4) the measured temperatures at various depths are shown by the line *A–A*. In this graph depths are indicated by the vertical scale at left and temperatures by the horizontal scale at top. *B–B* is a calculated graph showing vapor pressures (in kilograms per square centimeter, shown by the scale at bottom) of saturated water vapor at the measured temperatures. The graph *C–C* shows the calculated pressures of the hydrostatic column of water (in kilograms per square centimeter) at the various depths. Allowance has been made for the decrease in density of water with rising temperature (as measured), and the pressure of the barometric column at the elevation of 7,300 feet has been added. Some discussion of the significance of these data in the mechanism of geysers will be given in a later section.

THE GENERAL CHARACTER OF METAMORPHIC CHANGES

When microscopic examination of the core samples was begun there was little clue as to what was to be expected in the way of

metamorphic modifications, and preliminary examinations did not promise much information of value. Both in the sedimentary series from the upper part of the drill hole and in the lava flows occurring

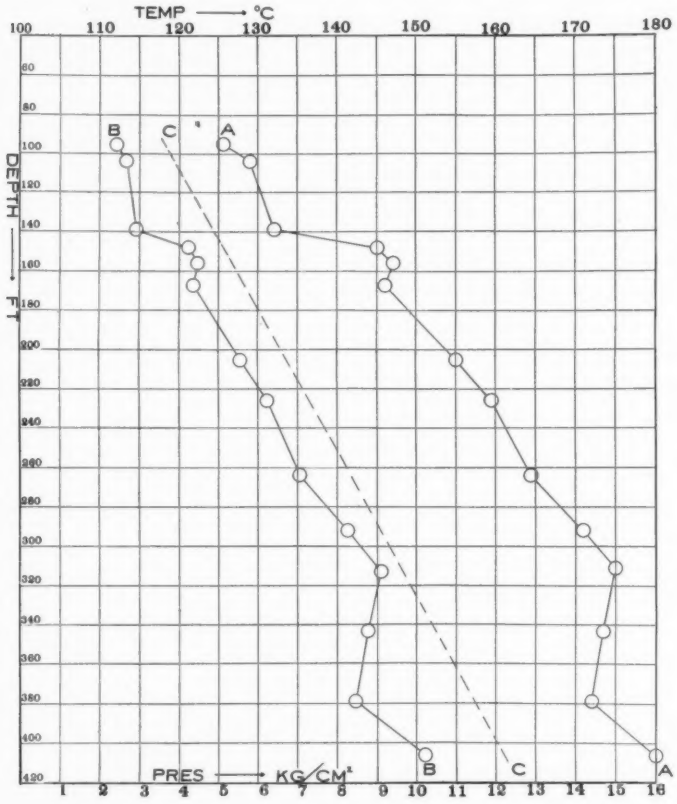

FIG. 4.—Graphs of temperatures and pressures in drill hole of Upper Geyser Basin. *AA* = measured temperatures, *BB* = vapor pressures of saturated water vapor at the measured temperatures, *CC* = hydrostatic pressures of a water column at the various depths. (1 kg./cm.² = 14.22 lbs./sq. in.)

at greater depth, phenocrysts of quartz, plagioclase, and orthoclase were readily recognized, and indicated that the whole series had orig-

inally been of rhyolitic or dacitic composition. In most specimens, however, the groundmass was of fine grain, made up of aggregates of crystals that individually had dimensions of only a few hundredths of a millimeter or even less.

Certain portions of the cores, however, showed somewhat larger crystals in the groundmass, and in some specimens cross-cutting veins made up wholly of secondary minerals of moderate size appeared; but on the whole it was seldom that the secondary minerals were of satisfactory development. By a rather tedious application of optical methods and analytical determinations, with occasional X-ray assistance kindly given by E. Posnjak and H. S. Roberts, the character of the secondary minerals and the nature of the metamorphic changes they represented were established.

In a broad way, the trend of metamorphism and the chemical processes involved are of simple character, but in detail there is considerable complexity. In the following pages petrographic descriptions of the rocks will be given, together with chemical analyses, and their significance as evidence upon which conclusions have been reached will be pointed out. Other features also will be described, which are not directly related to the main problem but are of interest in themselves. In order that the bearing of the data upon conclusions may be evident during the course of the descriptions, and not obscured by details or by the minor departures from the essentially straightforward course of major processes, a brief preliminary summary of the most important results will be given.

The sediments found in the upper part of the bore hole in the Upper Basin resulted from the deposition of fragmental material derived from the disintegration of rhyolites and dacites. Apparently mechanical processes had been chiefly responsible for the disintegration, and there had been little change of composition brought about by slow weathering and leaching.

In the lavas found below the sediments the original composition was similar, but tended more toward a dacitic composition. Their phenocrysts are chiefly sodic plagioclases; pyroxene phenocrysts are not rare but quartz and orthoclase are in minor amount.

In the whole series, vitrophyres greatly prevailed over other types, and abundant relics of perlitic structures are preserved.

During metamorphism the phenocrysts were little attacked, but the groundmass has been almost wholly metasomatized. This has been accomplished by the upward percolation of hot waters carrying potassium salts and silica in solution. Silica was deposited and potash was substituted for the soda and lime of the plagioclase molecules of the groundmass. By this process the close approximation to the unit ratio of alkalis plus lime to alumina that characterized the original rocks was preserved.

The mineralogical result in most cases was that the groundmass was replaced by orthoclase and quartz, and these are by far the most common secondary minerals. In the upper part of the series, by a reversal of the soda-potash interchange, the zeolites analcite and heulandite were deposited and may even now be replacing secondary orthoclase.

CHARACTERISTICS OF THE CHIEF SECONDARY MINERALS

Occasionally crystals of secondary quartz, several millimeters in length, may be observed with a binocular magnifier, projecting into cavities. In some of the thin sections of cores smaller crystals of quartz, with good development of pyramids at each end of the prisms, or as spindle-shaped forms, may be seen as a lining of pore spaces. More commonly quartz occurs as a mosaic of fine or minute grains intergrown with orthoclase or other minerals. Usually it is easily recognized as a chief constituent.

With regard to secondary orthoclase, several problems presented themselves in the course of the investigation, and had to be worked out in order to establish with certainty what had occurred, and its significance. First, it had to be determined that the abundant minute grains intergrown with other minerals in the groundmass mosaic really were orthoclase; second, that no other potash-bearing mineral of importance was present; third, the composition (potash-soda content) was to be determined; and fourth, it was necessary to show that the secondary orthoclase replaced in molar proportions the soda-lime feldspars.

As this study of the secondary orthoclase was applied to the Norris cores as well as to those of the Upper Basin, and as the results are closely interrelated, it seems best to summarize the two together at

this point, though the main descriptions of the two sets of cores are later taken up separately.

In several specimens crystals of secondary orthoclase of large enough size to be separated and examined were present, and these were studied microscopically and by X-rays. In other cores little vugs or spongy masses yielded crystals of orthoclase (together with silica minerals) of larger size than the general run in the groundmass mosaic. Likewise in the narrow veins that cut the rocks orthoclase is a chief constituent.

Some of the diagnostic optical properties in all these occurrences were measurable. Partial analysis of such material determined the potash-soda ratio, and this could be correlated with optical properties. The minerals of these veins or little masses could be seen to spread into the adjacent fine-grained mosaic with no change except diminution of size. The very small water content found in this material and in the general core substance excluded a number of possible potash minerals. Numerous complete analyses of cores were made, and in typical specimens the very small solubility of the constituent minerals under treatment with HCl was established. All these (with minor exceptions that will be noted) confirmed the presence of a large amount of orthoclase and the absence of other potash minerals.

The basis for the most important conclusion—that the potash now present is largely due to molar replacement of soda and lime in the original rocks—rests chiefly on two forms of evidence: the phenocrysts that have been preserved unchanged are not appropriate to rocks of the present composition, yet the unit ratio of strong bases to alumina that characterized the rocks originally has been maintained; and a certain specimen of core that has escaped alteration and that has the properties that much of the original material should have had (as indicated by phenocrysts) serves as a basis for comparison with the metamorphosed rocks.

OPTICAL PROPERTIES OF ORTHOCLASE AND PLAGIOCLASE IN THE CORES, AS RELATED TO COMPOSITION

As is well known, orthoclase feldspars usually contain some of the albite molecule in solid solution, and in some instances this amounts to a large percentage. As to the effect of this admixture upon the

optical properties of the mineral we find conflicting data, and, in general, there is much uncertainty regarding the mutual relations of the several forms of orthoclase with their varying optical properties. The material of the Yellowstone cores is not well suited for a complete study of optical properties, but it was possible to obtain a considerable amount of useful information. Both phenocrysts and groundmass material were studied, and informative results were obtained. Plagioclase phenocrysts are also important as witnesses of the original composition of the lavas, and results obtained from them will be cited here.

The sediments of the Upper Basin were doubtless derived from nearby sources, but were of somewhat mixed material. Phenocrysts of quartz, orthoclase, and plagioclase are associated in them, and indicate derivation from rocks varying from rhyolitic to dacitic. In the underlying lavas, quartz and orthoclase phenocrysts are rare, and plagioclase is the common feldspar. Pyroxene phenocrysts are not infrequent.

In Norris Basin the lavas of the cores contain abundant phenocrysts of quartz and orthoclase. Originally they were doubtless typical rhyolites.

The principal optical determination of value was that of indexes of refraction of grains of tested orientation in liquids that had been freshly standardized. When necessary, correction for temperature was made.

The secondary orthoclase was obtained from little nests of relatively coarse crystals that are present in some of the porous specimens, or from the intersecting veins. Care was taken to avoid inclusion of primary phenocrysts. Many specimens were tested, and, altogether, several hundred determinations of indexes were made on grains whose optical orientations had been ascertained.

With plagioclases, the soda-lime ratio is shown closely enough by the indexes, but with orthoclases large variations in the relative amounts of soda and potash produce very little change in indexes, or, according to some published data, practically none. Whatever may be the situation in general, the orthoclases in the Yellowstone cores show a small but determinable and consistent difference of indexes between those in which the amount of soda is insignificant and those in which there is a large amount of soda.

In core specimen Y.P. 750, from a depth of 221 feet in Norris Basin, phenocrysts were picked out of coarsely crushed powder under a binocular magnifier, and a partial analysis was carried out. No attempt was made to separate orthoclase and quartz, as the presence of quartz does not affect the ratio of bases, which is the important matter. A little pyrite was present as films along cleavage planes of feldspar, and this accounts for the iron found. Sulfur was not determined. The analysis is shown in Table I.

TABLE I

Y.P. 750A. PARTIAL ANALYSIS OF PHENOCRYSTS

	Weight, per Cent	Molar Ratio
SiO_2	80.63	.....
Al_2O_3	9.98	0.098
Fe_2O_3 (total Fe)	0.48	.....
CaO	0.28	.005
MgO	Tr	.....
Na_2O	2.72	.043
K_2O	4.61	0.049
BaO	0.03	.....
Ignition loss	0.38	.....
Total	99.11	.....

Molar $\frac{K_2O+Na_2O+CaO}{Al_2O_3}=\frac{.097}{.098}$.

Molar composition $=Or_{50.5}Ab_{44.3}An_{5.2}$.

This, then, is an orthoclase in which the molar quantity of Na_2O is not much less than that of K_2O (0.043 Na_2O to 0.049 K_2O), and very little CaO is present. The indexes, obtained by averaging numerous determinations, are as follows:

$$\alpha=1.523_3\,; \qquad \beta=1.528_4\,; \qquad \gamma=1.529_8\,.$$

Six measurements of $2V$ gave values ranging from 35° to 42°, average 39°, $r>v$. Sign negative.

From the same crushed sample, grains of metamorphosed groundmass were picked out for the determination of alkalis. This material was white, sugary, and appeared to be completely metamorphosed. No plagioclase was present, to which Na_2O could be referred. This is true of the secondary feldspar of all cores.

Feldspar content of material analyzed = 49.29 per cent Or, 30.34 per cent Ab by weight (calculated from Table II).

Molar ratios in the orthoclase crystals alone = 60.27 Or, 39.73 Ab.

By treating the powder of Y.P. 750 with boiling, dilute HCl ($1HCl:4H_2O$) for 15 minutes, only 0.08 per cent Na_2O and 0.14 per cent K_2O was extracted.

From the immediately adjacent core, Y.P. 749, little crusts of secondary minerals, consisting of a mixture of orthoclase, quartz,

TABLE II

ALKALIS IN Y.P. 750B

	Weight, per Cent	Molar Ratio
Na_2O	3.58	0.058
K_2O	8.33	0.088

and tridymite, were obtained, and indexes of orthoclase determined. This material is believed to be identical in composition with that whose analysis is given just above.

$$\alpha = 1.524 \pm .001 \; ; \quad \gamma = 1.530 \pm .001 \; .$$

From the results obtained with Y.P. 750 and Y.P. 749 it appears that the orthoclase in both the original phenocrysts and in the metamorphosed groundmass contains much soda, that the ratio of K_2O to Na_2O is somewhat higher in the groundmass than in the phenocrysts, that the indexes are essentially the same in the two, and that they are higher than those generally given for normal potash orthoclase.

Many confirmatory determinations of indexes were made on Norris cores from various depths. For secondary orthoclase, the results were as follows:

$$\alpha = 1.522 \text{ to } 1.524 \pm .001 \; ;$$
$$\beta = 1.528 \pm .001 \; ;$$
$$\gamma = 1.529 \text{ to } 1.530 \pm .001 \; .$$

Several measurements of axial angle of this secondary orthoclase were made. The values for $2V$ vary from 33° to 52°. The variation may be due to subparallel growths that distort the figure, but with β and γ so nearly identical a little change in either would cause a real change in $2V$.

Other measurements of phenocrysts of orthoclase gave the following values for indexes:

$$\alpha = 1.523 \pm .001; \quad \beta = 1.527 \text{ to } 1.528 \pm .001;$$
$$\gamma = 1.528 \text{ to } 1.530 \pm .001; \quad 2V = 32° \text{ to } 42°.$$

In the cores of the Upper Basin, phenocrysts of orthoclase are present in much less quantity than plagioclase. It was not practicable to separate the two feldspars for analysis, but determinations of indexes were made. The orthoclase phenocrysts had practically the same indexes as those given for the phenocrysts and secondary orthoclase of Norris Basin. On the other hand, most of the *secondary* orthoclase from this hole had definitely lower indexes and, correspondingly, the composition of practically pure potash feldspar. The data are as follows:

In Y.P. 289 (depth of 274 ft.) a narrow white vein consists wholly of quartz and orthoclase. Determination of alkalis gave K_2O 8.94 per cent, Na_2O 0.24 per cent, equivalent to 52.90 per cent Or, and 2.03 per cent Ab by weight in the vein material. The indexes of the orthoclase crystals were

$$\alpha = 1.518 \pm .001; \quad \beta = 1.523 \pm .001; \quad \gamma = 1.524 \pm .001.$$

In many other specimens from the Upper Basin secondary orthoclase appearing in veins or crusts or as scattered crystals gave $\alpha = 1.518 \pm .001$, $\beta = 1.523 \pm .001$, $\gamma = 1.523$ to $1.524 \pm .001$. These data for the Upper Basin are all consistent, but in two specimens from the upper part of the hole, secondary orthoclase was found in association with analcite, and has the slightly higher indexes of the high-soda variety found in the Norris cores. From its association with analcite it may well be of high soda content.

It appears from these data that in the Yellowstone rocks orthoclase high in soda may be definitely distinguished from strictly pot-

ash orthoclase by slightly higher indexes of refraction. As additional evidence, the indexes of sodic orthoclase from several specimens of lithophysae from Obsidian Cliff were measured. An analysis of such material is cited by Iddings,[4] and shows 4.16 per cent Na_2O and 3.96 per cent K_2O. The indexes are:

$$\alpha = 1.523 \text{ to } 1.524 \pm .001 \; ; \quad \gamma = 1.530 \pm .001 .$$

Still further results were obtained from a specimen of rhyolite from the cliffs of Madison Valley. No analysis has been made of this rock, but as all published analyses of unaltered Yellowstone rhyolites show high soda, the orthoclase of this specimen may fairly be supposed to be of this character. The rock carries phenocrysts of orthoclase and quartz in a porous, miarolitic matrix, in which apparently autogenous pneumatolytic action has formed small crystals of orthoclase and quartz, and little spangles of hematite. The phenocrysts of orthoclase gave

$$\alpha = 1.523 \pm .001 \; ; \quad \beta = 1.527 \pm .001 \; ; \quad \gamma = 1.528 \pm .001 .$$

Crystals from the miarolitic cavities of the matrix gave

$$\alpha = 1.523 \pm .001 \; ; \quad \beta = 1.528 \pm .001 \; ; \quad \gamma = 1.530 \pm .001 .$$

By means of the exhaustive chemical and optical studies cited it has been possible to draw the following conclusions, which could hardly have been reached with confidence on the basis of the rather conflicting data given in the literature: that for these rocks at least a high soda content of both primary and secondary orthoclase is accompanied by an increase in the indexes of refraction,[5] and that differences of composition of the secondary orthoclase formed under different environment are thus found to exist.

It remains to speak a few words regarding the plagioclase phenocrysts in the cores from the Upper Basin. It will be recalled that they are the principal phenocrysts in these lavas. Their composition

[4] J. P. Iddings, *U.S. Geol. Survey Monograph 32*, Part II (1899), p. 426.

[5] Winchell and Alling, who have made special studies of orthoclase, agree upon this, but other writers give discordant data.

is important as witnessing the character of the lavas before metamorphism. The crystals are somewhat zoned, and the indexes vary correspondingly, but the great bulk of the grains give consistent results

In Y.P. 289, the plagioclase phenocrysts give $\alpha = 1.542_6$; $\beta = 1.547_0$; $\gamma = 1.550_2$, indicating $Ab_{73}\,An_{27}$. This is the core in which the white vein material was found to consist of quartz and practically pure potash orthoclase. Phenocrysts of plagioclase from other cores of this hole have about the same indexes as Y.P. 289.

THE OCCURRENCE OF ZEOLITES

In the Upper Basin the zeolites, analcite and heulandite, were found in considerable quantity in cores from the upper part of the drill hole.

The analcite occurs in veins and as replacements of the groundmass. In veins many of the crystals show bright faces of 1–2 mm. dimensions. The mineral was identified by its crystallographic and optical properties, and the identification was confirmed by comparison of the X-ray pattern with that of known analcite, kindly made by H. S. Roberts.

Heulandite is present in abundance in some cores from the Upper Basin, but its vertical range is small. Its identification was difficult because of the exceeding minuteness of grains and their vague optical properties. After optical examination had failed to give much information, a chemical analysis was carried out on rather impure material separated from associated minerals (see analysis 215A in Table VI). Tests of water content, solubility in acid, and fusibility were also made. These suggested the mineral that Schaller has called "clinoptilolite,"[6] and on request he kindly supplied a specimen of his type mineral for X-ray comparison. This was carried out by H. S. Roberts, and the agreement of lines was definite. Later E. Posnjak found that the X-ray patterns of this mineral from Yellowstone Park, a certain mineral found by M. N. Bramlette in altered pyroclastics from California,[7] and heulandite from West Paterson,

[6] W. T. Schaller, *Amer. Mineral.*, Vol. VIII (1923), pp. 93–94; and Vol. XVII (1932), p. 128.

[7] M. N. Bramlette and E. Posnjak, "Zeolitic Alteration of Pyroclastics," *Amer. Mineral.*, Vol. XVIII (1933), p. 167.

New Jersey, were practically identical. Clinoptilolite appears to differ from heulandite chiefly in a substitution of part of the CaO in heulandite by K_2O and Na_2O in clinoptilolite, and perhaps a molar increase in SiO_2. The Yellowstone mineral appears to be of this "clinoptilolite" character. As the atomic space patterns are practically identical, it is doubtful if it should be regarded as a different species.[8]

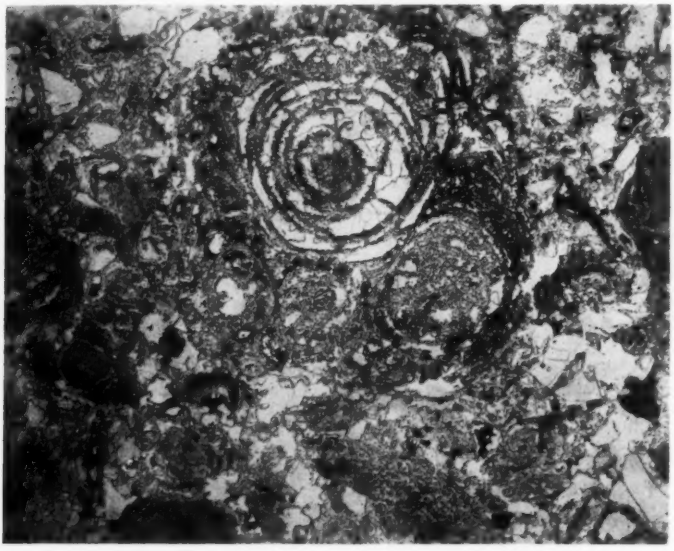

FIG. 5.—Photomicrograph showing relic structures due to replacement of perlitic glass. A few fragments of lithoidal rhyolite and quartz phenocrysts are present. The septa are formed of minute crystals of secondary quartz and orthoclase. Analcite is abundant, especially as a filling of cavities. Considerable goethite is present. Y.P. 232, from sedimentary series at depth of 138 feet in Upper Basin. ×16 diam. Plain light.

In the cores from the Upper Basin, the vertical range of heulandite is from 62 feet to 86 feet. It occurs as thin septa formed by deposition in perlitic cracks in obsidian, with subsequent removal of most of the obsidian (see Fig. 5); as films of tissue-paper thinness covering rhyolite pebbles; and as a replacement of the groundmass.

[8] In a recent conversation, Dr. Schaller has taken the same view.

MINERALOGY AND PETROGRAPHY OF CORES FROM THE UPPER BASIN

The drilling of the holes has given an unusual opportunity to study the results brought about by the action of thermal waters upon rock under known conditions. The results are different in some respects from anything that has been described heretofore.

The methods of study have been indicated in the preceding descriptions.

Much of the transformation has taken the form of metasomatic replacement of solid rock, but some of the mineral deposition was in open cracks or cavities. In some of the veins the size of grain is relatively coarse, but the individual crystals are seldom so large that they can be recognized with a hand lens.

For convenience of description the core specimens will be taken up by somewhat arbitrary groups. Under each group general features will be described, and specific details will be given. The significance of certain features will be discussed.

Cores from surface to 52 feet.—The specimens show a transition from a pure siliceous sinter at the surface to a gravel thoroughly cemented by silica. In the neighborhood of the drill site no surface deposition of sinter appears now to be taking place, and in many other localities in the Park surface deposits of silica are found where no waters now issue. The points of egress appear to shift from place to place. In the drill cores fissures are found that evidently at one time provided channels for rising waters but became filled by deposition of minerals, while other cracks have recently opened and only a film of minerals has been deposited on their walls.

The change from sinter to cemented gravel is gradual. At a depth of 7 feet there is a considerable amount of rhyolite pebbles, somewhat altered, together with glassy grains of quartz and orthoclase that represent loose phenocrysts. The sinter here contains finely crystalline quartz mixed with amorphous silica. There is no obvious means of determining whether the amorphous silica gradually crystallizes to quartz or whether each was deposited in the form now present.

At 20 feet pebbles predominate, but much siliceous sinter and secondary quartz are present. A staining test with a solution of the

organic dye, malachite green, gave negative results; therefore clay minerals are probably absent.

At 42 feet crystalline quartz appears in abundance, and the core shows numerous curious structures that, for a while, were very puzzling. Similar structures (Fig. 5) are present in many cores to the greatest depth reached. Study has shown beyond doubt that they represent a relic structure derived from perlitic obsidian. They reproduce faithfully the familiar spherical and polygonal cracks, and various stages of their formation have been found. The first step was deposition of new minerals in the perlitic cracks, followed by encroachment upon the adjacent glass. In many, the intervening glass was removed in solution before much had been replaced, and frail, curving septa are left, that resemble minute lithophysae.

In some of the cores this structure is so prevalent as to indicate that the original sands and gravels were composed mostly of disintegrated perlitic obsidian. Recent accumulations of mechanically disintegrated obsidian may be seen today as unconsolidated deposits in various places in the Upper Geyser Basin.

In the core from a depth of 42 feet the septa are formed of secondary quartz. A great many rhyolite pebbles are present. The whole aggregate is tightly cemented with secondary quartz.

Y.P. 208 (47 ft.) is a typical specimen of the rock at these levels. The thin section shows: (*a*) numerous fragments of rhyolite, some still showing typical flow structure in spite of considerable replacement by quartz; (*b*) original phenocrysts of quartz and orthoclase; (*c*) perlitic nodules and obsidian fragments, replaced by quartz; (*d*) almost unaltered fragments of the microcrystalline spherulites common in obsidians, with plumose or radiating structure; (*e*) much secondary quartz in grains and plumose crystals; (*f*) a small amount of fine-grained green chlorite; (*g*) a small amount of an isotropic mineral of low index, either opal or analcite.

In general make-up this specimen consists of pebbles and grains that were originally rhyolite, perlitic obsidian, and spherulites, together with grains of original quartz and orthoclase, set in a hard matrix representing the silicification of the finer groundmass. Several short, veinlike gashes are composed of a pale-blue translucent mineral that has the appearance of chalcedony. Under the micro-

scope it is almost cryptocrystalline, and has an index that suggests a mixture of quartz with opal or chalcedony. Narrow bands of slightly varying refractive index are visible with half-shade illumination. Similar material is present in adjacent cores. Some of these veins have central vugs lined with megascopically distinct crystals of quartz.

Cores from 57 feet to 121 feet (Y.P. 210 to Y.P. 228).—All the cores of this group consist of gravel or sand cemented by secondary products. In general, they are very firm, though many are at the same time porous. Small vugs or pores are lined with minute crystals, and narrow veins (2–3 mm. wide) of secondary minerals are present. To casual examination many of these cemented gravels resemble ordinary concrete mixtures in which pebbles are prominent.

Original phenocrysts derived from the disintegration of rhyolites occur, usually in broken fragments. Quartz and orthoclase are the most abundant, plagioclase is less common. Occasionally pyroxene crystals appear, and rarely zircon. The phenocrysts of quartz and feldspar lack evidence of replacement by secondary minerals.

Many of the pebbles consist of fragments of the ordinary nodular spherulites of rhyolites. They preserve their typical fine-grained, radiating, often plumose structure. Pebbles of altered lithoidal rhyolite are common.

The secondary minerals of the groundmass form fine-grained aggregates. Quartz and orthoclase are the chief products.

Obsidian, either as fragments with perlitic cracks or as loose perlitic nodules of the size of fish roe or a little larger, was originally abundant in the gravels. It is now represented by secondary minerals deposited along the cracks, where they form crystalline septa.

The sedimentary series as a whole (above 220 feet) is shown by its unaltered phenocrysts to have been derived from rhyolites of somewhat different character from the rhyolites or dacites *in situ* found below 220 feet. Apparently the sediments represent more siliceous and more potassic lavas than do the deeper flows.

In Y.P. 211 (62 feet) heulandite and secondary orthoclase make their first appearance. It was at this depth that steam pressure was encountered. The heulandite becomes a chief constituent at a little greater depth, and disappears at 86 feet. Secondary orthoclase con-

tinues to the bottom of the hole, and becomes, next to quartz, the chief metasomatic mineral.

Analcite appears in Y.P. 217 (86 ft.) (Fig. 6), and is present in many cores. It extends beyond this group to a depth of 192 feet and appears again in a single specimen at 216 feet. In some specimens it forms distinct veinlets and frequently it occurs as an impregnation

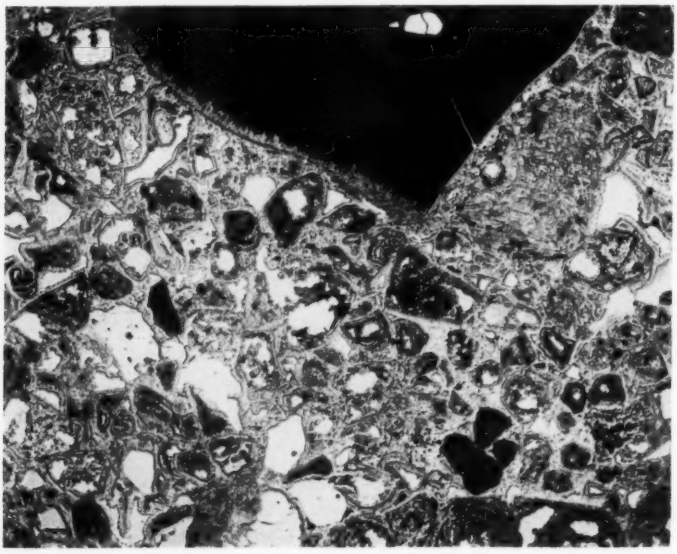

FIG. 6.—Photomicrograph showing fragments of obsidian and lithoidal rhyolite, replaced by minute crystals of heulandite, quartz, orthoclase, and analcite. Phenocrysts of quartz, orthoclase, and plagioclase are present. In the upper portion is a large fragment of a spherulite, which appears nearly opaque, but actually has the typical structure of rhyolitic spherulites. Specimen Y.P. 217, from sedimentary series of Upper Basin, at depth of 86 feet. ×16 diam. Plain light.

and replacement of the general mass. Crystals from Y.P. 226 were tested by X-ray methods by Mr. Roberts. This gave certainty to its determination, but its identification was seldom doubtful. In veins, its crystals give mirror-like reflections. Even where it occurs as a replacement mineral it tends to assume a crystal outline.

Its distribution is sporadic. In some specimens in which veinlets

of it may be seen, thin sections of the core do not show it. In cores between 192 feet and 216 feet it was not found, but it appeared again at 216 feet.

Quartz is the most common secondary mineral, not only of this group but of all cores (with a few exceptions) to the bottom of the hole. It often appears as anhedral, equant grains, but in porous cores little crystals with perfectly developed faces of prism and pyramid project into cavities. In larger cavities megascopic crystals are recognizable. Plumose growths also occur, some of which have the normal properties of quartz and others resemble chalcedony.

Opal occurs in several specimens of this group. In Y.P. 214 (78 ft.) it assumes veinlike forms. Its color here is variable: milky white, brown, pale blue, or colorless and transparent. The index varies from less than 1.460 to considerably more. The last undoubted opal is seen in Y.P. 217 (86 ft.).

A little chlorite is almost always present throughout the series. Some of it is disseminated, but it commonly collects in little spherulites or nodules. The nodules themselves are aggregates of extremely minute crystals, too small for the determination of specific optical properties.

Tests of rock powder of core specimens with a solution of malachite green sometimes showed the spherulites of chlorite to be stained to a deeper color, but clay minerals are not recognizable, and if they are present at all the amount is very small.

Calcite comes in sporadically in this group. In some thin sections it forms small, irregular patches, which plainly replace other minerals. It occurs also in association with quartz, orthoclase, etc., in veinlets, as megascopically visible thin tablets flattened parallel to the base and with the edges truncated by rhombohedral forms (argentine). Some of these crystals are of comparatively large size: a centimeter or more in their two greatest dimensions, but only 0.5–2 millimeters thick.

Within the small vertical range in which heulandite is present it forms a large part of the core material. A typical specimen in which it occurs is seen to contain sharply bounded dark grains and pebbles of rhyolite, together with phenocrysts of quartz and orthoclase and some plagioclase, set in a nearly white matrix, which, in the thin

section, is seen to consist of heulandite, quartz, and orthoclase. Both in the hand specimen and in the section innumerable little shelly structures of the secondary minerals indicate that the original material of the matrix was chiefly glassy obsidian with perlitic cracks. Heulandite, quartz, and orthoclase were deposited in these cracks, and the intervening glass was removed.

In Y.P. 218 (92 ft.) heulandite has disappeared. The core consists of pebbles of altered rhyolite, together with phenocrysts of quartz, orthoclase, and plagioclase, in a fine-grained matrix, which is now composed of quartz and orthoclase and disseminated chlorite, but which was originally a sandy filling composed principally of obsidian. A large white aggregate of secondary minerals, with drusy cavities, is found to be a mixture of quartz, analcite, and calcite. The quartz is in plumose form; associated with it is material which has the properties of chalcedony. Analcite covers chalcedony, and in turn it is covered with calcite.

A few minute grains of yellow sulphide are present.

In the next core (92 ft.) pebbles are almost absent. In this and succeeding cores through Y.P. 223 (97 ft.) the secondary minerals are essentially quartz, orthoclase, and analcite.

Sandy material with few pebbles again appears from 97 feet to 116 feet. Secondary analcite is present here in large amount.

In Y.P. 226 (116 ft.) the indurated sand is cut by a narrow vein. This contains open cavities lined with small crystals of quartz in broadly plumose growths overlain with brilliant little crystals of analcite, and these are covered in places with aggregates of platy calcite. This has the optical properties of ordinary calcite, and effervesces briskly in cold, dilute HCl.

In Y.P. 227 (120 ft.) the rock was originally composed essentially of small grains of perlitic obsidian and a few pebbles. In the thin section phenocrysts of quartz and orthoclase are seen, several plagioclases, and one pyroxene. The matrix has been replaced by quartz and orthoclase and a little calcite. Analysis of the specimen shows 8.72 per cent of K_2O and only 0.98 per cent of Na_2O (see Table VI).

A special test for soluble alkalis was made on this specimen. Four grams of the powder was boiled with 25 cc. concentrated HCl + 100 cc. H_2O for 15 minutes, and alkalis were determined in the solution.

The results were 0.06 per cent Na_2O and 0.20 per cent K_2O (analysis Y.P. 227B in Table VI). Although the core had been found to contain 8.72 per cent K_2O and 0.98 per cent Na_2O, very little K_2O and practically no Na_2O are soluble in HCl. The presence of zeolites and such minerals as alunite is thus excluded. The specimen only 4 feet above this and the specimen 1 foot below show veins of analcite, but evidently this mineral is absent here. This illustrates the sporadic distribution of analcite.

In Y.P. 228 (121 ft.), one end of the specimen is covered with a thin, irregular vein of analcite in little crystals, but no analcite appears in the thin section. Instead, there are numerous crystals of secondary quartz, minute but well developed, projecting into pore spaces, and much secondary orthoclase, some of it in lath-shaped crystals that are apparently contemporaneous with quartz. Brown patches of tiny, acicular crystals of high index and high birefringence occur in this specimen and many others, in places closely associated with pyroxene phenocrysts, and there evidently an alteration product. The quantity is small, and few optical properties can be determined, but it is believed to be goethite.

Orthoclase is present in the analcite vein in this specimen, and this is one of the occurrences previously mentioned where secondary orthoclase in the cores from this site has the higher indexes of the soda-bearing variety.

In some respects the group of cores just described is the most interesting of this series, especially in the variety of the metasomatic minerals. Opal, chalcedony, quartz, calcite, heulandite, analcite, orthoclase, chlorite, and goethite are present. Opal is not abundant and is found only in the upper levels. Heulandite is the chief mineral in a few cores and then drops out abruptly. Analcite occurs especially in veins, but it is also an important replacement mineral. It continues to somewhat deeper levels. Calcite is found chiefly in veins, but not in large amount anywhere. Quartz and orthoclase become abundant, and continue to the lowest depths reached by the hole.

An important question is whether the temperature and other conditions that now prevail represent the conditions under which these minerals were formed or whether they are a survival from pre-

vious conditions. It is believed that both are partly true, and that the minerals are a mixed assemblage.

Among the secondary minerals, quartz and orthoclase are found in great quantity in the upper levels, and continue to the bottom of the hole. It can hardly be doubted that they are stable at the lower levels, but there may be a question with regard to their presence near the surface. Right at the surface there is a great amount of amorphous silica (sinter), but no quartz. Opal and quartz are found together at shallow depths, and below only quartz. Opal, however, is not a truly stable mineral. Though it is frequently deposited where there is a rapid change of conditions in silica-bearing solutions, and may persist indefinitely, yet the tendency to change is from opal into quartz. The presence of opal near the surface cannot, therefore, be regarded as evidence that quartz is unstable there.

With orthoclase, however, the evidence points to another conclusion. The aluminosilicate minerals most prominent at shallow depths are heulandite and analcite, and analcite continues in little veins down to 216 feet. It seems significant, also, that in two instances (in cores from 121 and 132 ft.), where it was possible to determine the indexes of orthoclase directly associated with analcite, they were of the higher values that indicate soda-bearing orthoclase. The interpretation that seems most probable is that the strictly potash orthoclase found in the upper levels represents a survival from previous conditions, and that the process by which potash was substituted for soda, formerly active in the upper levels and still going on at greater depth, is no longer operative in the upper levels. It might at first be supposed, in explanation of this, that the waters that now reach the upper levels have reacted as far as they could at lower levels and are simply in a state of equilibrium with regard to the potash-soda interchange, and that the zeolites are a recrystallization in hydrated form of residual albite; but when we examine the analysis of Y.P. 215 (Table VI), which is a typical heulandite core, we find such high soda and low potash relative to the typical orthoclase cores that we conclude that there is an actual reversal of the process.

There are, however, certain minor irregularities in mineral occurrences in these cores that probably do not represent a general change of conditions but only irregularities of movement of solutions along

channels. In veins in two of them there is a definite sequence of deposition observable: quartz-analcite-calcite; and this might seem to indicate that quartz, first, and analcite, later, ceased to be deposited. This is doubtless true as regards the particular places where this sequence is observed, but simultaneous deposition of these same minerals may have continued at points in immediate proximity.

In the core from 92 feet a cavity has been filled with quartz, analcite, and calcite, deposited in the order given, whereas another cavity in the same specimen, only half an inch away, is lined with megascopic crystals of quartz, with nothing deposited on them. Evidently the conditions were not so uniform that deposition of analcite and calcite at one point necessitated their simultaneous deposition at adjacent points.

These rocks show varying degrees of porosity, and many narrow fissures intersect them. In an open channel, as compared with a finely porous rock, movement of solutions is facilitated and the duration of contact with unaltered rock is shorter. Temperature is a trifle higher or lower, and concentration of dissolved material greater or less, depending upon directness of communication with regions below from which heat and emanations are contributed, or with cooler regions from which meteoric waters enter. For these reasons it is apparent that conditions on which precipitation of a certain mineral is dependent are not uniform at a given level throughout the mass. Movements of solutions from pores into channels or from channels into pores produce changes affecting deposition, and the gradual filling of a channel may determine what minerals are deposited on its walls. The paragenetic succession observed may thus be partly dependent on local causes. It is necessary to consider results in the mass before it can be concluded that a certain local sequence implies a general change of conditions.

Cores from 122 feet to 219 feet (Y.P. 229 to Y.P. 263).—This group, like the preceding, was originally a series of sediments. Gravel is present in considerable amount, but the quantity is less than in the upper levels. Sands predominate. In the lowermost 25 feet, lying immediately above the dacite basement, much of the material was a very fine sand or a silt. Mingled with this are many isolated pebbles larger than the diameter of the core. The presence of these large

pebbles in silty sediments suggests the dropping of pebbles from floating ice.

All the material is much altered. It is solidly cemented, yet in many of the originally sandy cores it is very porous. In this group of cores, as in those from all levels, narrow veins of secondary minerals occur. Some of them contain crystal-lined vugs. Their dips vary from nearly vertical to nearly horizontal. The sharpness of the breaks shows that fracturing was subsequent to induration, and in some instances the small amount of mineral deposit on the walls indicates recent formation.

Narrow fractures, such as these, dipping at various angles, associated with little or no crushing, and apparently formed at intervals over a prolonged period, seem most likely to have been brought about by local readjustments. We are inclined to ascribe them to changes in the volume of the batholith that we believe to underlie the region and to be the source of the thermal phenomena.

This body may be decreasing in volume, both by loss of heat and by loss of substance through emanations. The resulting contraction seems a likely cause of fracturing in its own mass and in the roof.

Y.P. 230 (128 ft.) is typical of the uppermost specimens of this group. It is composed of firmly cemented sand. The structure is very porous, and the porosity is partly due to the many little spherical shells resulting from the solution of perlites. One end of the core is cut off by a fracture that dips about 20°. This appears to be recent, for the crust that covers the surface is hardly more than a film of minutely drusy crystals, though there are a few larger crystals of platy calcite. A thin section of the specimen shows many fragments of quartz phenocrysts, some of orthoclase and plagioclase; spherulites, large and small; perlitic relics, and grains of altered rhyolite. The chief secondary minerals are coarse analcite, quartz, and orthoclase, together with a little calcite, chlorite, and goethite.

Y.P. 231 (132 ft.) is similar. It is cut by a narrow (1–1½ mm.) vein, irregular but sharply bounded, dipping about 65°, and containing small vugs. The vein consists mostly of intergrown quartz, orthoclase, and analcite. The vugs are lined with analcite, which covers quartz and orthoclase, and a single platy crystal of calcite is seen.

Y.P. 234 (146 ft.) is composed of small pebbles firmly cemented in a cellular white matrix, whose pores glisten with minute crystals. The core is cut off diagonally by a steeply dipping fracture, which is covered with a film of intergrown quartz and orthoclase. The thin section shows numerous pores lined with quartz and orthoclase, likewise intergrown. The orthoclase is in crystals which are unusually large for these rocks, though of only 0.03–0.04 mm. dimensions. The quartz is in little prisms terminated by pyramids.

Y.P. 235A (147 ft.) represents a core section through a single large pebble of altered rhyolite. Its mass is two or three thousand times that of the common run of pebbles in the cores.

Y.P. 237–Y.P. 242 (168–183 ft.) are fine, even-grained sands, firmly coherent. Shelly forms representing perlites are numerous. The chief secondary minerals are quartz and orthoclase in all, and analcite in several.

Y.P. 244 (186 ft.) carries two large pebbles of rhyolite separated by a band of silt.

The succeeding cores through Y.P. 259 (214 ft.) show mixtures and alternations of large rhyolite pebbles with silts or very fine sands. The silty cores are stratified in a somewhat irregular manner. The large pebbles are commonly of a reddish brown color. This may be due to weathering prior to incorporation in the sediments, or it may have been produced by previous fumarole or hot spring action. If the hypothesis is correct that the deposition of this series of sediments followed shortly after glacial action had scoured away surface rocks, it is hardly likely that thoroughly weathered material was available. Several of these cores carry sharply defined, flaky inclusions of a reddish-brown to brick-red color (Fig. 7), set in the midst of greenish-gray to olive-green silts. These flakes have the appearance of the crusts that form around "mud-pots" or "paint-pots" in various places in the geyser basins. Those present in the thin sections are made up of fine, stratified material, of a reddish color, cut off abruptly. Ordinary weathering seldom produces such brilliantly colored mud flakes under the climatic conditions of the Yellowstone. This evidence seems to support the idea of thermal activity having been in progress before the beginning of deposition of the series of sedi-

ments. This supplied pigmented mud flakes and oxidized rhyolite pebbles, which were rafted out and dropped among the silts.

Y.P. 248 (192 ft.) shows unusual features. Megascopically it appears to be a cemented sand. In the thin section many fragments of phenocrysts and spherulites of the ordinary sort are seen. Mingled with them are fragments that have the typical shapes of shards of

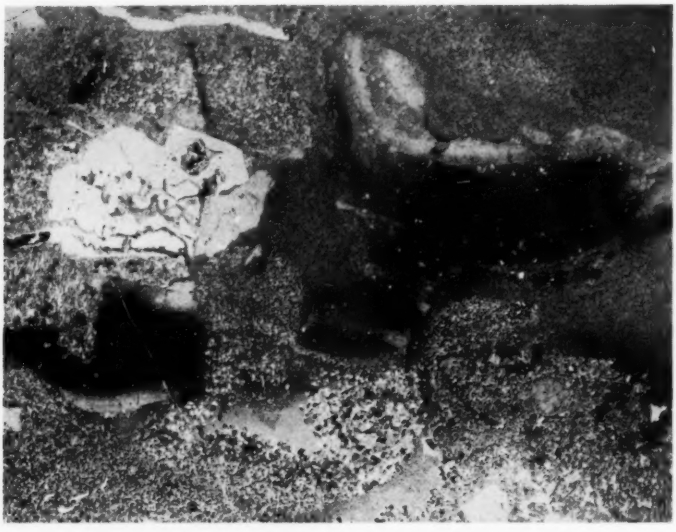

FIG. 7.—Photomicrograph of mud flakes (stained with iron), surrounded by grains of sand and silt, from Y.P. 256, at a depth of 206 feet in Upper Basin. (Large white area on left is a plagioclase phenocryst.) ×16 diam. Plain light.

volcanic glass, altered to aggregates of exceedingly minute crystals, of axiolitic arrangement. These shards of glass are a constituent not recognized in other cores.

In the next core (192 ft.) analcite appears for the last time until it reappears in small quantity in Y.P. 260 (216 ft.).

The principal secondary minerals of this group of cores are quartz, orthoclase, and analcite; others of less abundance are hematite, goethite, and calcite. Another mineral, which has not been identi-

fied, occurs in very small amount here and throughout a considerable range. It forms little pockets of very slender cottony fibers.

The fragments of phenocrysts in these sediments are still chiefly quartz. Orthoclase is common but plagioclase is becoming more abundant. Notwithstanding the thorough metasomatism of the matrix, structural features, such as sedimentary stratification, are well preserved. It is somewhat surprising to find that a specimen which has all the appearance, megascopically, of a banded clay is of rock-like hardness and that it is made up chiefly of quartz and orthoclase.

Y.P. 260 (216 ft.) is an olive-green indurated silt of this sort, cut by a barely perceptible vein of glassy analcite, which connects small cavities lined with brilliant little crystals of analcite. This is the last observed appearance of this mineral.

Y.P. 261 (217 ft.) is a similar olive-green silt. The thin section shows replacement by fine-grained aggregates of quartz and orthoclase, which are cut by many little veins of coarser crystals, some of which are composed almost wholly of quartz and others of intergrown quartz and orthoclase. Certain crystal-lined cavities show orthoclase resting on quartz, but this is believed due to local conditions rather than to be indicative of a general succession. The large amount of orthoclase is indicated by the 9.68 per cent of potash of the analysis (Table VI). There is an unusual amount of the tiny crystals that are believed to be goethite, and the analysis shows an exceptional amount of Fe_2O_3.

Calculation of molar ratios of alkalis plus lime to alumina in this analysis gives 1.027. If this originally silty sediment had been a clay produced by ordinary weathering, much of the alkalis would have been leached out, and it is very doubtful if the later metamorphosing solutions would have restored alkali to unite with uncombined alumina. This would have required a different sort of chemical process from the *base interchange* that has prevailed. We infer that the silt had been little acted upon by weathering, but was probably undecomposed rock flour produced by glacial scour.

Y.P. 263 (219 ft.) is the last of the sedimentary specimens. It was originally a silt. The thin section shows replacement by quartz and orthoclase. Several large, irregular cavities in the core, which ap-

pear to have been made by solution, contain quartz crystals 1–2 mm. long.

Cores from 220 feet to 291 feet (Y.P. 264 to Y.P. 298).—At 220 feet the drill hole entered the lava flows that underlie the sediments. All of the cores of this group (except Y.P. 292) consist of igneous rock, more or less metamorphosed. Many cores still contain much glass, and in others perlitic structures survive after complete alteration. Phenocrysts are fairly numerous. Of these, plagioclase crystals are most abundant, and pyroxene and magnetite are common, but quartz and orthoclase are rare.

Fortunately, one specimen, previously referred to (Y.P. 279, from 260 ft.), is almost unaltered. It doubtless represents part of an "island" that the thermal waters have not penetrated. An analysis made of this specimen gives a valuable standard by which the changes of composition during metamorphism may be judged. Probably the original composition of these lavas varied somewhat, but in many of the cores the phenocrysts have the same character as in Y.P. 279, and the Ab:An ratio of the plagioclase phenocrysts is nearly constant. It is believed, therefore, that the composition of Y.P. 279 is representative. A description of this may serve as a basis for comparison and discussion of others.

The core is a dark-brown obsidian or pitchstone, with perlitic cracks. It has a glassy to resinous luster and a "crisp" appearance. The thin section (Fig. 8) shows a uniform, pale-brown, isotropic glass containing short, hairlike microlites in sparse distribution. Perlitic structure is perfectly developed. A few of the cracks contain films of a pale-green decomposition product, probably chloritic, in insignificant amount. The glass has an index of 1.491. In other cores above and below this we see various stages of alteration of similar glasses, and the incipient stages are easily detectable. In this, such effects are insignificant; one seldom sees a volcanic glass of fresher appearance. It is believed that in spite of the large amount of water present, this specimen preserves its original composition.

Phenocrysts, some in glomeroporphyritic groups, are of small amount in proportion to glass. They consist of twinned feldspar, pyroxene, and magnetite. The feldspars are biaxial, negative, and are faintly zoned. The general indexes are $\alpha = 1.541$; $\beta = 1.546$; $\gamma =$

1.550, corresponding to Ab74 An26. The pyroxene phenocrysts deserve special notice. Most of those in the thin section give biaxial, positive figures of moderate angle, but two give a uniaxial, positive cross, and one gives a biaxial, negative figure. The uniaxial crystals are intergrown with others that give the usual biaxial figure. All look normal. In immersion liquids the indexes of the two varieties

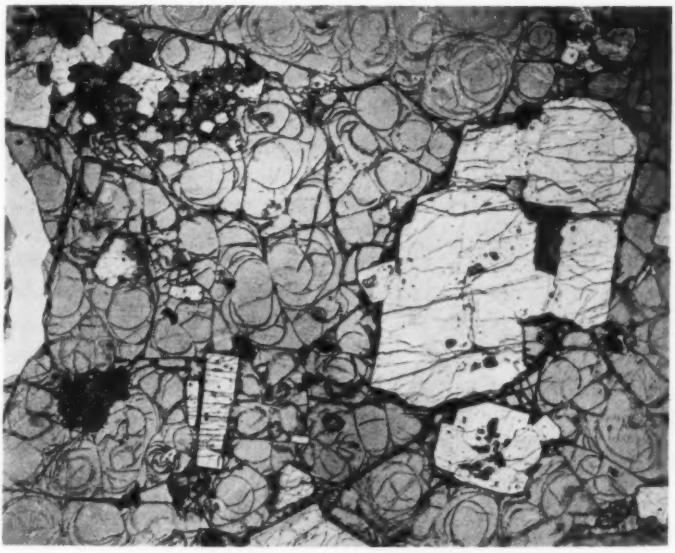

FIG. 8.—Photomicrograph of Y.P. 279, an unaltered vitrophyre with perlitic structure, from a depth of 260 feet in the Upper Basin. The specimen is part of an unreplaced "island" among the flows of dacite. The photograph shows phenocrysts of plagioclase and pyroxene in glass. ×16 diam. Plain light.

are found to be not very different. The lowest index is a little below 1.725 and the highest a little above 1.740.

The analysis of this specimen (Table III) shows 6.31 per cent of water and considerable chlorine and sulfur. Carbon dioxide is absent or negligible.

The composition is shown in column 1 of Table III, and recalculation with volatiles omitted is given in column 2. In column 3 there

is added for comparison the only analysis of a rhyolite from Upper Geyser Basin quoted by Iddings.[9] He says of it that it might properly be classed as dacite.

The volatile-free composition of Y.P. 279 has considerable resemblance to the analysis quoted by Iddings, but it has more lime and less potash.

TABLE III

UNALTERED DACITES OF UPPER GEYSER BASIN

	1	2	3
SiO_2	67.09	71.71	70.92
Al_2O_3	13.08	13.97	13.24
P_2O_5	0.08	0.08	0.18
TiO_2	0.47	0.50	0.16
Fe_2O_3	1.60	1.71	3.54
FeO	1.94	2.08	0.66
CaO	1.89	2.02	1.42
MgO	0.53	0.59	0.23
MnO	0.05	0.05	0.14
Na_2O	3.97	4.24	4.28
K_2O	2.75	2.94	4.25
BaO	0.09	0.10	.....
Cr_2O_3	0.00	0.00	.....
ZrO_2	0.02	0.02	.....
H_2O+	4.23	.....	0.57
H_2O-	2.08	.....	
CO_2	0.01	.....	.....
Cl	0.05	.....	.....
SO_3	0.15	.....	.....
Total	100.08	100.01	99.59

1 = Y.P. 279.
2 = Y.P. 279 with volatiles omitted.
3 = Iddings' specimen.

In many other cores of this group phenocrysts of uniaxial, positive pyroxenes (pigeonite) were found. The plagioclase phenocrysts are consistently of basic oligoclase (about Ab70 An30). The rocks of the cores were probably not a single flow, but the uniformity of phenocrysts indicates close similarity of composition.

The effect of slight hydrothermal metamorphism of a rock of the composition of Y.P. 279 is shown in Y.P. 267 (225 ft.). The specimen is similar to Y.P. 279, but devitrification has begun along per-

[9] Iddings, *op. cit.*

litic cracks, and each of them is occupied by minute crystals, whereas the intervening glass is little attacked.

The composition, shown by analysis (Table VI), is almost a duplicate of that of Y.P. 279, except for an increase of K_2O and decrease of Na_2O. The increase of the one is the molar equivalent of the loss of the other, and we attribute this to substitution by thermal waters. By comparison of these two specimens it is possible to distinguish clearly the first effect of hydrothermal alteration.

A number of the specimens still consist largely of volcanic glass. In others, the only surviving evidence is the little spherical structures originating from perlitic cracks. A few of the completely metamorphosed specimens give the impression of having been formed from flow-banded felsites.

In many specimens of this group the groundmass has been completely recrystallized, and consists now of a fine-grained intergrowth of the usual quartz and orthoclase, with a little goethite, but in some specimens the glass has undergone a different type of alteration. It has not been found possible to identify satisfactorily the fine-grained products of this second type, but they are evidently of different character.

This latter type of alteration does not seem to fit into the otherwise consistent scheme. Considerable effort has been made to determine the nature of the minerals and the significance of the difference. Some information has been obtained, but several features remain unexplained.

Y.P. 273 (238 ft.) is a typical specimen. Under the binocular magnifier the core is seen to have been originally a perlitic glass, with feldspar and pyroxene phenocrysts. The altered rock is of greenish-gray to light-brown color and of dull, earthy luster. The appearance is much like that produced by ordinary weathering.

The thin section shows the glassy groundmass to have been wholly altered and to consist now of several secondary products, not all of which are satisfactorily determinable. We find the following:

1. Minutely crystalline, bright-green nodules and elongated forms along cracks, and as an alteration of pyroxene. The mineral is probably a chlorite.

2. Many small aggregates of a yellowish-brown, finely acicular

mineral, of high index, high birefringence, and positive elongation. Much of this occurs as an alteration of pyroxene. From the optical properties and from the chemical results that will be given, it is believed to be goethite.

3. Numerous large, fibrous spherulites and plumose growths. The extinction is parallel to elongation, which is γ. The index of refraction is low and the birefringence moderate. It suggests a zeolite. This mineral has a characteristic appearance in the thin section, but in rock powder it cannot be distinguished in the general mixture.

4. Minutely crystalline material, obviously a mixture of several minerals; too fine-grained and too much intergrown to be identifiable.

The chemical work upon this specimen consisted, first, of an analysis of the usual sort, whose results are given in Table VI of analyses; and, second, an extraction test based upon the fact that treatment of the powder with boiling dilute HCl, followed by digestion of the residue with a hot solution of Na_2CO_3, should have little effect upon the primary phenocrysts or upon the secondary quartz and orthoclase that are the common metamorphic products in the cores, but should decompose more or less completely a number of minerals, including most zeolites.[10]

The results of this procedure were as shown in Table IV, in percentages of the whole.

Common minerals that might be decomposed and taken into solution (wholly or partly) by the acid and alkali treatment are zeolites, chlorite, opal, and hydrated ferric oxide. Feldspars would be affected

[10] An outline of the procedure is as follows: The 100-mesh powder was boiled for 15 minutes with $1HCl:5H_2O$, and the solution filtered. The filtrate was evaporated to dryness and the small amount of soluble SiO_2 was thus separated. The filtrate from this separation was divided into two parts. In one the bases were determined in the usual manner. In the other, all the bases except alkalis and lime were precipitated by addition of freshly ignited CaO, and the CaO was precipitated by ammonium carbonate and oxalate. Alkalis were then determined in the filtrate.

The residue from the first treatment of the rock powder with HCl was washed into a platinum basin, using about 65 cc. of water. 6.50 grams $Na_2CO_3H_2O$ was added, and the solution was kept near boiling temperature for 10 minutes, agitating frequently. This presumably extracted the gelatinous SiO_2 left from the decomposition of silicates by HCl, and opaline silica originally present. The resulting solution was filtered and acidified, and SiO_2 determined as usual.

little or not at all by the dilute acid that was used, as is shown by the results obtained with Y.P. 227B (see Table VI) and Y.P. 750C (Table XII). It is evident, therefore, that in the mineral aggregate there is much material that is different from the quartz-orthoclase mixture that is the usual result of alteration. Alkalis, lime, and alumina are among the chief constituents extracted, and this points to zeolites. The large amount of water shown by the main analysis is in accord. The soluble ferric oxide may be ascribed to goethite. The preponderant substance extracted is silica, and it seems as if a part of this (in excess of that required for zeolites) should be referred to opal.

A large amount of alkalis, lime, alumina, and silica was left undecomposed. This naturally suggests feldspars and quartz. It looks,

TABLE IV

EXTRACTION ANALYSIS OF Y.P. 273 B

SiO_2	13.16	MgO	0.11
Al_2O_3, P_2O_5	3.14	Na_2O	1.61
TiO_2	0.14	K_2O	0.44
Fe_2O_3	1.57	Total extracted	21.17
CaO	1.00		

therefore, as if two sorts of alteration had acted upon the rock, one giving the quartz-orthoclase mixture that is the normal result in most rocks at this depth, and the other producing zeolites and opal.

The chemical investigation has strengthened the inference to which the microscopic examination led, that a considerable portion of the metamorphic minerals is of different character and different chemical behavior from those in the cores above and below them. In comparison with the unaltered pitchstone (Y.P. 279) there has been only a small increase of K_2O and no increase of SiO_2. It seems that although these rocks are thoroughly altered, only a part of the alteration is due to the potash-silica-bearing waters that produced the typical alteration in the others.

It is difficult to advance a reason why rocks of different metamorphic character should be intercalated in the series here, but the possibility may be suggested that the specimen is part of an "island"

that lacked main channels of circulation. The penetration and movement of solutions, therefore, were so slow that little foreign material was contributed, and recrystallization of the original constituents of the glass was effected in practically stagnant solutions.

In this part of the drill section there are a great many cores in which much glass is still present, and many others showing the earthy type of alteration. Those of typical quartz-orthoclase character are rather few.

Several of the earthy cores were tested, in powder, with malachite green dye. In most cases the fine particles slowly absorbed the stain and became deep green. This probably indicates zeolites rather than clays. In addition to the minerals described under Y.P. 273, little nests of the cottony fibers mentioned previously are seen in a number, and the thin sections of several show an isotropic mineral of very low index which looks like opal.

In the analyzed specimen of this type (Y.P. 273) there was not the large increase of K_2O over Na_2O that is characteristic of the normal hydrothermal alteration. This was believed to be their most significant chemical characteristic. In three other specimens of the earthy type alkalis were determined, with similar results.

TABLE V

ALKALI DETERMINATIONS

(In Percentages)

	Y.P. 281 262 ft.	Y.P. 285 271 ft.	Y.P. 290 276 ft.
Na_2O	2.90	2.54	2.66
K_2O	3.80	3.71	3.16

These ratios are very unlike those found in specimens of the quartz-orthoclase type of alteration, such as Y.P. 227, 261, 325, and 334.

Dehydration tests at a low red heat showed these specimens to contain between 6 and 8 per cent water (by loss), thus differing from the normal type in this respect also.

The last appearance of a rock of the earthy type is at 287 feet.

The quartz-orthoclase type of alteration in these dacites gives not very different mineralogical results from those described for the overlying sediments. Quartz and orthoclase greatly exceed all other minerals. A little calcite appears in a number of thin sections, and patches of goethite are present in most, but chlorite has almost disappeared. Perlitic structures survive in the manner described.

Many of the specimens of metamorphosed lava contain angular inclusions, up to a centimeter in size, of darker color, similarly metamorphosed but usually spongy. They are interpreted as representing fragments of glass in a lithoidal dacite.

Of the original phenocrysts, plagioclase is much more common than quartz and orthoclase. Optical tests show it to vary from $Ab_{67}An_{33}$ to $Ab_{74}An_{26}$. Pyroxene commonly shows partial alteration to goethite, magnetite is preserved, and a few zircons have been noted.

White veins of secondary quartz and orthoclase, with a little calcite, are numerous, but only a few millimeters wide. In Y.P. 288 (274 ft.), such a crust rests upon metasomatized dacite, with a sharp demarcation between vein and wall. At one place it has been broken away by later movement, and several small flakes of the wall rock have been detached. A film of minute crystals has been deposited subsequently over the whole, covering the fractured surfaces and cementing the detached flakes in position. Crystals of secondary orthoclase from this vein were examined by H. S. Roberts, by X-rays, and their structure identified with that of known orthoclase.

In Y.P. 289 two veins of quartz and orthoclase intersect. A partial analysis of this material was made, as noted on an earlier page. It gave K_2O 8.94 per cent, Na_2O 0.24 per cent, H_2O 0.41 per cent. These figures correspond to about 53 per cent Or and 2 per cent Ab by weight.

In most of these narrow white veins quartz is somewhat predominant, but locally orthoclase is in excess. Considerable calcite is present in some.

Ordinarily, in cores in which veins are present, the adjacent rock is hard and dense. It is of gray or greenish-gray color, and there is a sharp boundary between vein filling and wall rock, but when a section across the contact is examined microscopically not much miner-

alogical difference is found. The wall rock has been almost completely metasomatized and it also consists essentially of quartz and orthoclase, though usually in smaller crystals than in the vein. The original phenocrysts of feldspars, quartz (if present), and magnetite have seldom been much attacked, but pyroxene is somewhat altered to finely acicular goethite. Chlorite is seldom distinguishable in this group.

All but one of these cores are quite certainly metasomatized lavas, but Y.P. 292 (284 ft.) is evidently made up of cemented sand, consisting originally of small rhyolite or dacite grains, spherulites, perlites, and many angular fragments of quartz, orthoclase, and plagioclase distributed in sedimentary fashion. Metamorphism has produced the usual quartz-orthoclase secondary minerals. This sediment must have been deposited between flows of lava, as the cores immediately above and below are altered dacites. Presumably it has no relation to the series of sediments above the dacites, for it is separated from them by more than 60 feet of lava flows. Originally it may have been covered by several hundred feet of lava.

Cores from 293 feet to 400 feet (Y.P. 299 to Y.P. 334).—All these specimens show the quartz-orthoclase replacement of dacites or rhyolites, and present very little that is different from what has been described.

The phenocrysts are predominantly of plagioclase (approximately $Ab_{70}An_{30}$), pyroxene, and magnetite, with an occasional zircon crystal. A few have quartz and orthoclase. The metasomatic minerals, which completely replace the original groundmass, are quartz and orthoclase in great excess. Subordinate goethite is present, and small spots of segregated hematite occur. There is a very little sulfide ore. Calcite is present sparsely but in large plates.

Some of the cores of this group are dense and hard, others are granular and porous. The brown specks and spongy areas previously described are common. Indications of perlitic structure are shown by some, and others have a streaked or even a ribbed appearance, as if they had been banded flows. A few look as if they had been pumiceous.

Many specimens are cut by white veins of quartz and orthoclase, with a little calcite. They dip at various angles. Several show the

appearance described above, as if later movement had broken away part of the crust, exposing the underlying rock, and then a new deposit of crystals had been laid down.

In Y.P. 314 (321 ft.) the thin section has the appearance of welded glass fragments, such as Iddings has described and figured.[11] When viewed with crossed nicols the streaked groundmass is seen to have been replaced by quartz and orthoclase.

ANALYTICAL RESULTS AND THEIR SIGNIFICANCE

In Table VI the results of complete and partial analyses of cores from the Upper Basin are assembled.

A summary of the character of material analyzed follows:

Y.P. 215A = portion of Y.P. 215 obtained by crushing the sample gently and sifting through 100-mesh screen. Represents impure heulandite, which is very friable and was thus separated from denser material. The analysis was made in preliminary effort to identify the mineral.

Y.P. 215B = more compact portion of Y.P. 215, left as residue on 100-mesh screen. Consists mostly of pebbles of altered rhyolite, phenocrysts of quartz and feldspar, and opal.

Y.P. 215 = calculated combination of Y.P. 215A and Y.P. 215B according to proportions of the two.

Y.P. 227 and Y.P. 261 = metamorphosed sediments. Y.P. 227 was originally a perlitic sand; Y.P. 261 was a fine silt. Both have been replaced by quartz and orthoclase.

Y.P. 267 = dacitic glass with a small amount of metasomatic minerals along perlitic cracks. The composition is almost a duplicate of Y.P. 279 except for decrease of Na_2O and increase of K_2O.

Y.P. 273 = altered dacitic pitchstone. Earthy type of alteration.

Y.P. 279 = unaltered dacitic pitchstone. Regarded as representative of original composition of many of the core specimens.

Y.P. 325 and Y.P. 334 = completely metamorphosed dacite. Normal type of alteration.

Y.P. 281, Y.P. 285, and Y.P. 290 = altered perlitic glass (earthy type of alteration). Alkalis were determined to show dissimilarity of this type to quartz-orthoclase type.

[11] Iddings, *op. cit.*, pp. 403–6.

TABLE VI

COMPLETE AND PARTIAL ANALYSES OF UPPER BASIN DRILL CORES

	Y.P. 215A (82 ft.)	Y.P. 215B (82 ft.)	Y.P. 215 (82 ft.)	Y.P. 227 (120 ft.)	Y.P. 261 (217 ft.)	Y.P. 267 (225 ft.)	Y.P. 273 (238 ft.)	Y.P. 279 (260 ft.)	Y.P. 325 (375 ft.)	Y.P. 334 (400 ft.)	Y.P. 281 (262 ft.)	Y.P. 285 (271 ft.)	Y.P. 290 (276 ft.)	Y.P. 289 (274 ft.)	Y.P. 227B (120 ft.)	Y.P. 273B (238 ft.)
SiO_2	67.61	74.85	70.07	75.36	67.18	66.65	66.51	67.09	77.36	78.35	.....	.....	.....	.....	.....	13.16
Al_2O_3	12.33	11.76	12.13	12.02	15.26	13.07	12.93	13.08	10.73	9.97	.....	.....	.....	.....	.....	3.14
P_2O_5	0.04	0.05	0.05	0.04	0.13			0.08	0.06	0.05	.....	.....	.....	.....	.....	
TiO_2	0.14	0.22	0.16	0.15	0.42	0.47	0.44	0.47	0.32	0.35	.....	.....	.....	.....	.....	0.14
Fe_2O_3	0.75	0.90	0.81	0.83	2.13	1.68	2.08	1.60	1.18	1.44	.....	.....	.....	.....	.....	1.57
FeO	0.60	0.65	0.62	0.46	1.20	1.88	1.22	1.94	0.88	0.95	.....	.....	.....	.....	.....	
MgO	0.13	0.10	0.12	0.06	0.62	0.59	0.43	0.53	0.42	0.33	.....	.....	.....	.....	.....	0.11
CaO	1.43	0.87	1.24	0.80	1.45	1.92	2.62	1.89	1.09	1.19	.....	.....	.....	.....	.....	1.00
MnO	n.d.	0.01	.....	0.02	0.05	n.d.	n.d.	0.05	0.04	0.01	.....	.....	.....	.....	.....	.....
Na_2O	3.00	2.88	2.96	0.98	1.52	3.38	2.58	3.97	1.02	0.96	2.90	2.54	2.66	0.24	0.06	1.61
K_2O	2.66	4.07	3.14	8.72	9.68	3.59	3.26	2.75	6.16	5.68	3.80	3.71	3.16	8.94	0.20	0.44
BaO	0.05	n.d.	.....	n.d.	n.d.	0.12	n.d.	0.09	n.d.	n.d.	.....	.....	.....	.....	.....	.....
Cr_2O_3	0.00	n.d.	.....	n.d.	n.d.	0.00	n.d.	0.00	n.d.	n.d.	.....	.....	.....	.....	.....	.....
ZrO_2	n.d.	n.d.	.....	n.d.	n.d.	0.06	n.d.	0.02	n.d.	n.d.	.....	.....	.....	.....	.....	.....
H_2O+	5.58	2.67	4.59	0.36	0.63	3.86	4.50	4.23	0.65	0.50	.....	.....	.....	.....	.....	.....
H_2O-	5.20	1.12	3.81	0.07	0.26	2.39	3.40	2.08	0.18	0.21	.....	.....	.....	.....	.....	.....
CO_2	0.29	0.13	0.20	0.40	0.05	0.08	0.07	0.01	0.50	0.04	.....	.....	.....	.....	.....	.....
Cl	0.008	n.d.	.....	n.d.	n.d.	n.d.	n.d.	0.05	n.d.	n.d.	.....	.....	.....	.....	.....	.....
SO_3	0.20	n.d.	.....	n.d.	n.d.	0.13	n.d.	0.15	n.d.	n.d.	.....	.....	.....	.....	.....	.....
Total	100.02	100.28	99.90	100.27	100.58	99.87	100.04	100.08	100.59	100.03	.....	.....	.....	.....	.....	.....

Y.P. 289 = quartz-orthoclase vein cutting metamorphosed dacite.

Y.P. 227B = alkalis extracted by boiling the powder of Y.P. 227 in 25 cc. HCl + 100 cc. H_2O for 15 minutes. Results show that alkalis are almost entirely in insoluble minerals.

Y.P. 273B = material extracted from Y.P. 273 by the acid and alkali treatment described in the text (p. 266).

In several places reference has been made to these analyses to illustrate features described. It has been shown from petrographic data that the most important processes in the alteration of the rocks were the deposition of silica and the substitution of potash for soda. With the analyses before us, further discussion of processes may be given.

In the original rhyolites and dacites of this series of rocks (and in such rocks in general) the gross composition is equivalent to a mixture of quartz and feldspar principally, and other minerals to a minor degree. These compounds may be present as crystals or they may be occult in a glass. In feldspars the molar ratio $(Na_2O+K_2O+CaO)/Al_2O_3$ is equal to unity. Therefore, an approximately 1:1 ratio characterizes rhyolites and dacites. Some departure from unity is to be expected because of the presence of a small amount of pyroxene or other ferromagnesian mineral, in which the ratio of alkalis plus lime to alumina may vary in either direction from 1:1. In the analyses of fresh rhyolites and dacites of Yellowstone made years ago by Gooch and Whitfield, we find that few show much departure of their molar ratios from unity.

In the unaltered core that was analyzed (Y.P. 279) the ratio is found to be almost exactly 1:1 (0.992), and this specimen is undoubtedly fairly representative of much of the original material, though in portions of the rock series the ratio may have been a little greater or a little less.

If, during metamorphism, there had been differential leaching of such a character as to remove alkalis or lime on the one hand or alumina on the other, the ratio would have been correspondingly altered. The actual results are shown in Table VII in the column headed $(Na_2O+K_2O+CaO)/Al_2O_3$.

From these results it appears that in the altered specimens, both sediments and lavas, the 1:1 ratio has been closely maintained. The

departures from unity are not greater than they may have been in the original rocks.

This, however, does not tell quite the whole story. In many of the rocks calcite is present in small quantity. It is visible in the cores and is indicated by the CO_2 of the analyses. The lime in this combination cannot fairly be paired with alumina, except to show that lime has not been removed from the system in disproportionate amount. If the molar quantity of lime was originally united with alumina (whether in feldspars or in ferromagnesian minerals is immaterial), then its later mineralogical separation from alumina, as calcite, means that an equivalent amount of alumina has been set

TABLE VII

CORES FROM UPPER BASIN: MOLAR RATIOS

Sample	Depth (Feet)	Al_2O_3	CaO	Na_2O	K_2O	$\frac{Na_2O+K_2O+CaO}{Al_2O_3}$	$\frac{Na_2O+K_2O+CaO-CO_2}{Al_2O_3}$
215....	82	0.119	0.022	0.048	0.033	0.866	0.832
227....	120	.118	.014	.016	.093	1.042	0.966
261....	217	.150	.026	.025	.103	1.027	1.020
267....	225	.128	.034	.055	.038	0.992	0.977
273....	238	.127	.046	.042	.035	0.969	0.953
279....	260	.128	.034	.064	.029	0.992	0.992
325....	375	.105	.019	.016	.065	0.952	0.848
334....	400	0.098	0.021	0.016	0.060	0.990	0.980

free. The quantity is indicated by the differences between the molar ratios in the column headed $(Na_2O+K_2O+CaO)/Al_2O_3$ and those in the column headed $(Na_2O+K_2O+CaO-CO_2)/Al_2O_3$.

The small excess of alumina that is to be accounted for may be present as some clay mineral that has not been detected in the finely crystalline groundmass of the cores, but a more probable explanation is that it has gone to form chlorite from the break-up of pyroxene.

It will be seen later, when the Norris cores are described, that there alkalis and lime have been leached out to a greater degree than in the cores of the Upper Basin, and carried away completely, and clay minerals have been formed.

Analyses of the cores support the microscopic evidence that the newly formed minerals are almost wholly of such character as to pre-

serve the 1:1 ratio of strong bases to alumina, and they supplement in a gratifying manner and help to interpret the results obtained by Allen in his investigation of the hot spring waters of this basin. These waters show great preponderance of sodium over potassium and carry very small amounts of other bases.[12] The two most important *rock constituents* in the issuing waters are silica and the sodium ion. These, in comparison with all other ingredients, are present in much more than their proper relative amounts.

As for the sodium ion, the petrographic and chemical investigation of the cores gives convincing evidence that though the rocks have lost sodium, the process has been chiefly the replacement of sodium in the aluminosilicates by potassium supplied by waters rising from greater depths.

The loss of sodium has therefore been made up by its equivalent of potassium. This would not result in an increase of silica, yet there has been an increase of this constituent also. This is brought out most clearly by comparing the molar ratio of silica to the sum of the other non-volatile constituents in the unaltered pitchstone (Y.P. 279) with the corresponding ratios in the altered rocks. This specimen of pitchstone is representative of the original dacitic lavas below 220 feet. The sediments above 220 feet were of more rhyolitic composition, with quartz phenocrysts, but probably not as siliceous as they now are. Their molar ratios of silica to other constituents are also given in Table VIII, but the comparison with the unaltered rocks is not so direct. It appears, however, that among the altered rocks the dacitic lavas that were originally less siliceous than the sediments now have the higher ratios of silica to other constituents, and all have a decidedly higher ratio than the unaltered specimen.

Thus we find that though the issuing waters bring up silica in great relative excess over all other constituents except sodium, yet the rocks as far as the drill hole has explored are richer in silica than originally. Evidently waters rising from greater depths have brought in silica as well as potassium, and the silica in the issuing waters has

[12] Allen and Day find reason to believe that thermal waters in many other regions of the world may owe their character to similar reactions. E. T. Allen and Arthur L. Day, "Hot Springs of the Yellowstone National Park," *Proc. Fifth Pacif. Sci. Congress*, Vol. III (1934), p. 2278.

not been leached from the rocks that we are able to study. The amount brought to the surface is the excess that escaped deposition in the region explored.

Since potassium has replaced sodium in the albite molecules, it is natural to inquire whether an analogous replacement of calcium by potassium has occurred in the anorthite molecules. If there was such a tendency, free silica was readily available to supply the amount needed for orthoclase. There is evidence pointing to this replacement, though it is not so plain from the petrographic studies or from simple inspection of the analyses, as the quantities involved are

TABLE VIII

INCREASE OF SILICA DURING METAMORPHISM, SHOWN BY MOLAR RATIOS OF SILICA TO OTHER NON-VOLATILE CONSTITUENTS*

Y.P. 279 (unaltered dacitic pitchstone from 260 ft.)	3.61
Y.P. 215 (altered sediment from 82 ft.)	4.91
Y.P. 227 (altered sediment from 120 ft.)	4.91
Y.P. 325 (altered lava from 375 ft.)	5.44
Y.P. 334 (altered lava from 400 ft.)	5.68

* The rocks listed in this tabulation are those representing the typical potash-silica metamorphism that is by far the most prevalent process. Others analyzed to show certain special or anomalous features that have been described are not included.

smaller. In the original crystallization of the magmas, the feldspar phenocrysts, in accordance with the principles governing the crystallization of solid solutions, took up more than their relative proportion of anorthite; and the phenocrysts have been little altered subsequently. Much of the anorthite left in the groundmass, however, seems to have been replaced by orthoclase (except that a small amount formed calcite).

To show this, we compare the molar ratio $(K_2O+Na_2O)/CaO$ in the unaltered pitchstone with this ratio in the altered rocks. It would not do to compare K_2O alone with CaO, for the replacement of Na_2O by K_2O would itself increase this ratio.

Inspection of the figures in Table IX shows an increase of alkalis to lime. The comparison is most direct between unaltered and altered lavas (Y.P. 279 compared with Y.P. 325 and Y.P. 334). In the sediments (Y.P. 215 and Y.P. 227) the composition was more alkalic originally. One of the sediments (Y.P. 227), however, has a value

so much higher than is likely to have characterized the original composition that an increase is indicated. The other sediment (Y.P. 215) is one of those carrying a large amount of heulandite. On a previous page we saw reason to believe that in the near-surface rocks the potash-soda interchange is now proceeding in the reverse direction, and it appears from the low value for Y.P. 215 that the potash-lime interchange has likewise been reversed.

TABLE IX

MOLAR RATIOS $\frac{K_2O+Na_2O}{CaO}$

Y.P. 279	Y.P. 215	Y.P. 227	Y.P. 325	Y.P. 334
2.74	3.68	7.36	4.26	3.62

POROSITY OF THE ROCKS; FORMATION OF CAVITIES

No large open spaces were found in drilling, but the rocks are considerably fractured, and many have a large amount of pore space through which hydrothermal solutions easily move. The pore space may be even greater than is indicated in the cores recovered, for the recovered cores are the densest portions of the rocks passed through, and the less coherent material lost in drilling was a large proportion of the whole.

The most important chemical results were a replacement of soda and lime by potash and an increase of silica, but the very small quantity of other bases brought to the surface by the waters should not be neglected. This might represent a loss from the rocks that the drill hole penetrated, leaving them more porous, or, like silica and potassium, it might have been derived from greater depths.

Many of the cores are highly porous, but the original materials might have possessed this property in equal degree, or might have developed it by the shrinkage attendant upon the crystallization of glassy obsidian. There is direct evidence of solution in many of the cores, but others are more dense than they are likely to have been originally. It is not obvious whether, in the cores as a whole, there has been a net gain or loss of mass, and probably it is not of great importance to decide this. What is of much greater importance is

that they offer evidence that if certain conditions had been slightly different much material might have been removed in solution, not rapidly but over a period of centuries.

We have seen that not only were orthoclase, analcite, heulandite, and calcite formed as *replacements* of original minerals by the metasomatizing action of soluble salts (chlorides, carbonates, etc., of the alkalis), but that all these difficultly soluble minerals were taken into solution, to some extent, that they migrated rather freely, and were deposited in fissures and other open spaces. If these minerals can be carried from the wall rock into veins, presumably they could be carried farther and removed altogether if the solutions moved faster. The necessary condition is that slowly moving solutions, dispersed through a mass of wall rock, where they have picked up material, should enter less obstructed channels where the dissolved matter is carried away faster than it can be deposited. This condition might well be met in portions of the geyser basin where the discharge is greater. By this process certain favorable channels, probably of great irregularity, would be slowly dissolved out in somewhat similar fashion to that in which caverns in limestones are formed. Eventually some of them would reach the size and form required for the beginning of geyser action. When this occurred, the rate of solvent action would be accelerated, for in geyser eruptions the sudden discharge of large quantities of water carries out whatever mineral matter is in solution in all the channels and pore spaces that contribute to the discharge. Such a process, extending over a long period, seems well adapted to dissolve out and enlarge channels. We believe that a system of large and small, irregular channels, very different from the simple forms illustrated in textbook diagrams, is an essential feature of the geyser mechanism. The phenomena of geyser action require that reservoirs of this kind should be available to supply the discharge in its variable and complex manifestations, and we are unable to see any probable explanation of their formation other than by this process of solution.

COMPARISON WITH HYDROTHERMALLY METAMORPHOSED ROCKS IN OTHER REGIONS

In earlier pages it was stated that the alteration found in these rocks seems to be different in some respects from anything that has

been described. Nevertheless, it is doubtful if there actually is anything very different here from what occurs in other places.

If rocks that had been altered in the same manner were found in a region where the obvious surface evidences of hydrothermal action had been removed by erosion, and if they were examined petrographically in a routine manner, most of the lavas would probably be regarded as fairly normal dacites or rhyolites. The texture and general appearance of phenocrysts and groundmass are not so greatly at variance as to be impressive. If a more careful examination were made, by which it was determined that the groundmass consists almost wholly of quartz and pure potash feldspar, whereas the phenocrysts are composed of lime-soda feldspar and a considerable amount of pyroxene and iron ore, the inconsistency might be recognized. It might be observed, also, that in some of the rock sections the secondary quartz and orthoclase of the groundmass have an unusual appearance. They form euhedral little crystals instead of interlocking grains, especially where the rock is porous. If chemical analyses were made, the very high silica and potash and low soda and lime would emphasize the incongruous character of the phenocrysts, yet it might not be specially remarked. There may be many occurrences of similarly altered rocks in volcanic regions, which, in absence of complete data, have been regarded as normal lavas. Even analyses of such materials may have been accepted as representing original compositions.

A feature that might cause one to suspect that metamorphic processes had affected the composition of these rocks is the presence of quartz-orthoclase veins. Such veins indicate the passage of solutions rich in potassium and silica, and alteration of the general mass of adjacent rock might be regarded as probable.

In recent articles R. D. Terzaghi[13] has discussed the problem of the origin of potash-rich rocks from theoretical principles and from petrographic considerations. One of the conclusions is that there may be no potash-rich rocks among the products of primary crystallization, but that rocks of this character are likely to be due to secondary processes. The present writer is uncertain how generally this may hold, but the Yellowstone rocks illustrate the manner in

[13] R. D. Terzaghi, "The Origin of the Potash-Rich Rocks," *Amer. Jour. Sci.*, Vol. XXIX (1935), p. 369, and Vol. XXX (1935), p. 141.

which thorough potassic replacement may be accomplished without very noticeable change of appearance.

At times, geologists, having observed that the wall rocks associated with a plutonic mass do not show the evidences of alteration that are regarded as essential criteria of contact metamorphism, have concluded that the magma has had little or no effect upon them. It seems that more careful consideration is required.

SOURCE OF THE METAMORPHOSING SOLUTIONS

Some of the more important conclusions regarding the hydrothermal phenomena of Yellowstone Park, acquired by the Geophysical Laboratory during its several years of comprehensive investigation, have been summarized briefly by Allen and Day[14] and by Allen.[15]

These authors note that among the dissolved materials in the hot waters the chief acid radicals with which the alkalis are united are chloride, fluoride, and bicarbonate, with smaller amounts of sulfate (or sulfide), borate, and arsenate. None of these is present in more than traces in the neighboring rocks, but they or their closely related representatives are typical of volcanic and fumarolic emanations. They are therefore believed to issue in volatile form from an underlying magma.[16]

When these emanations leave the magma and travel through the

[14] E. T. Allen and Arthur L. Day, "Hot Springs of the Yellowstone National Park," *Proc. Fifth Pacif. Sci. Congress*, Vol. III (1934), p. 2275.

[15] E. T. Allen, "Geyser Basins and Igneous Emanations," *Econ. Geol.*, Vol. XXX (1935), p. 1.

E. T. Allen, "Neglected Factors in the Development of Thermal Springs," *Proc. Nat. Acad. Sci.*, Vol. XX, No. 6 (1934), p. 345.

[16] Distillation of volatiles from a plutonic magma through a porous roof is essentially the same process as escape of gases from a surface flow of lava open to the air. There is apparently a rather widely held idea that before gases escape in quantity from a plutonic magma, the magma must have advanced far in its cooling and crystallization, or even have passed into a supercritical condition. This is by no means necessary.

Among the volatiles driven off, some may be far above their critical temperatures, and others far below. Whatever their state, it does not imply a critical condition of the magma or of the volatiles dissolved in it.

The pressure that gases may exert in magmas that plainly have not approached a critical state is impressively demonstrated in volcanic explosions of immense violence. This is the same pressure that tends to drive gases out of a plutonic magma through the roof pores.

roof the tenuous vapors of the alkali halides are believed by Allen and Day to be swept along in a current of steam and carbon dioxide. At the lower limit of ground water they are taken into solution, and their further upward course is in the form of a dilute liquid solution.

A large part of the alkalis in the thermal waters is thus attributed by Allen and Day to sublimation from the magma, but another part of the alkalis and the silica is believed by them to be acquired in another manner.

A chief constituent of the magmatic emanations is CO_2. When this is dissolved in the heated aqueous solutions it attacks the silicate minerals of the wall rock and extracts bases, especially alkalis. The alkaline water then takes silica into colloidal solution in the manner commonly recognized. Evidence that alkalis are extracted from the rock in this manner is found in the fact that the waters carry considerably more of the alkalis than can be assigned to the halide radicals but is assignable to the bicarbonate radical. Since carbonates or bicarbonates of alkalis are very slightly volatile, the excess must have been acquired after the gases left the magma. This process accounts for the acquirement of silica and part of the alkalis.

The process thus formulated seems to be required to give a satisfactory explanation of the presence of alkali bicarbonates, although its effects are not directly observable in the cores from the Upper Basin. If alkalis and silica are acquired in this manner, alumina in excess should be left, but neither petrographic examination nor analyses have revealed it. Allen has suggested that the attack by CO_2 takes place below the depths explored. It seems as if this must be true as regards the Upper Basin, but a significant feature has been found in studies of the drill cores of Norris Basin, which were continued after the articles by Allen and Day were published, in search of evidence for the process suggested by them. Analyses of these cores show a considerable excess of alumina, and clay minerals are visible in fair amount. They seem to have been formed by a later process than that by which potash was substituted for soda in the groundmass, and to replace both groundmass and phenocrysts. This will be discussed later.

Those reactions of which there is more direct evidence in the rocks of the Upper Basin involve the transfer of an enormous amount of

potash from lower levels and its fixation at higher levels. In the typical metamorphosed specimens, Y.P. 227, Y.P. 261, Y.P. 325, and Y.P. 334, the average content of K_2O is 7.56 per cent, whereas in the unaltered pitchstone, Y.P. 279, it is only 2.75 per cent. A thickness of several hundred feet of rock has received an increase of probably 4.5–5.0 per cent of K_2O, and an unknown thickness of enriched material lies below.

THE BORE HOLE IN NORRIS BASIN: GEOLOGICAL SETTING AND NARRATIVE OF DRILLING

The Upper Geyser Basin had been chosen as a site for drilling because it represents a typical area of alkaline springs. The site in Norris Basin was selected as representing an area of acid sulfate waters. Allen's work had led him to believe that the acidity in such areas is a superficial condition, caused by the oxidation of H_2S, and that this characteristic gives way to the alkaline condition at a moderate depth.

After the completion of the boring in the Upper Geyser Basin in the fall of 1929, a hole was started in Norris Basin, but after it reached a depth of about 20 feet the lateness of the season and the rapid advance of cold weather made it advisable to postpone its completion to the next year.

In 1930, work was resumed about the middle of August and continued until the early part of October. This work, like that at Old Faithful, was done by Butler, and the same type of drilling machine was used.

At the Upper Basin the drilling had given very little trouble to one as experienced as Butler in drilling in hot ground, but at Norris very different conditions were met. Whereas at the Upper Basin the highest steam pressure measured was 57 pounds per square inch, at Norris a maximum pressure of more than 300 pounds was encountered. This is the highest pressure known to us for such projects, and it increased the difficulties enormously.

The site chosen was in an area of hot ground, not far from Congress Pool, and a little off the main highway. Nuphar Lake lies just to the northeast. Figure 9 shows the site, looking northward, with the drilling rig set up, and with a considerable area of Norris Basin

beyond. This photograph was taken at an early stage of operations, and it is evident that there are no steam vents in the immediate vicinity of the drill house. Later, the breaking-out of vents here was one of the disturbing phenomena (Fig. 10).

Unlike the site in the Upper Basin, there are no superficial gravel beds at this spot, except ordinary detritus, and little or no sinter deposits. Rock *in situ* comes close to the surface. Not far away to the

FIG. 9.—Drilling rig in Norris Basin at the beginning of operations. View looking northerly.

west, however, are the Ragged Hills, a broken ridge of roughly bedded but well-defined strata of silicified and cemented sands and gravels, rising 50–100 feet above the flat floor of Norris Basin. Other remnants of similar indurated sediments occur at several places in the vicinity. Their character suggests deposition by the wash of streams into a shallow lake, with subsequent silicification by hot springs, but the exact conditions of formation are unknown. They are evidently eroded remnants of a more widespread formation, and apparently are of considerable age.

In the immediate vicinity of the drill site the surface material covering the flat area is sand and gravel mixed with angular pieces of

rock. No rock outcrops appear for a considerable distance in any direction. Even the rising ground to the east shows none, though solid rock probably lies at a shallow depth, for the surface is pretty

FIG. 10.—Drilling rig in Norris Basin after steam had broken out in numerous places at the surface around the drill site.

well covered with angular fragments. These fragments are found around the drill site also, and consist of leached rhyolite, in which the original quartz phenocrysts are intact but the feldspar phenocrysts have wholly disappeared, leaving angular cavities of characteristic

form in the fine-grained groundmass. This kind of surface leaching has been found to be definitely associated with areas in which the thermal waters contain H_2SO_4. Allen has given much attention to the problem of the manner in which the H_2SO_4 has been formed, and has reached the conclusion, supported by a wealth of evidence, that it results from near-surface oxidation of H_2S in situations in which the topographic conformation permits only shallow penetration of meteoric waters. The data obtained from the drilling of the Norris hole support this view.

In the Norris hole the rock penetrated was evidently broken by fissures and joints. This is indicated by irregularities of pressures and temperatures. For instance, on September 9, at a depth of about 244 feet, a pressure reading in the morning gave 181 pounds, but the next day, after steam had been allowed to blow off for some time, the pressure had dropped to 88 pounds. On September 11, the reading in the morning was 89 pounds. During the day a fissure was passed through in drilling, and the pressure increased suddenly to 225 pounds. The next day the reading was 250 pounds.

The drill contractor became apprehensive that with such high pressures brought close to the surface the shattered rock would be blown out bodily and a great crater would be formed. Drilling was discontinued for several days, and a heavy mass of concrete, reinforced with iron tie-rods, was laid, covering a considerable area adjacent to the drill hole. Steam was allowed to blow off continually during this time, with no indication of diminution of pressure; in fact, a reading taken a few days later, before drilling was resumed, indicated 297½ pounds (above atmospheric pressure).

The temperature a few hours later was found to be 205°. The high temperature and pressure were undoubtedly due to steam coming up through fissures from a lower region. According to calculations, the pressure was more than sufficient to lift the weight of a column of rock extending from the bottom of the hole to the surface. Apparently it was only because of the interlocking of the jointed masses that there was not an upheaval. When the thermometer was examined after the last measurement, the surface of the glass was found to be blackened and etched, though it had been wrapped in cloth and placed in an iron receptacle.

When an attempt was made to resume drilling, the feed water pumped down failed to reach the bottom. Some of it was forced out through the threaded joints of the pipes and the remainder was vaporized. This was remedied with difficulty.

These great pressures and high temperatures rendered every phase of the work difficult and slowed down progress very greatly. After 20 feet more had been drilled a succession of new difficulties brought operations to a final stop.

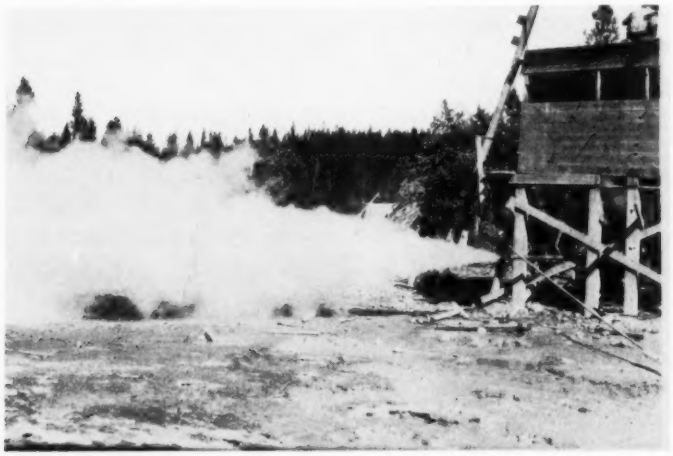

FIG. 11.—Escape of high pressure steam, Norris hole

The pressure, which had ranged from 297 pounds to 250 pounds, dropped rather suddenly to 125 pounds. This looked as if the steam were finding new outlets, and in a short time it was evident that it was breaking through to the surface. Blowouts of steam, accompanied by crumbling and ejection of ground, appeared about 25 feet to the south, and the openings increased rapidly in size, but before they reached an alarming condition, caving stopped. A multitude of other vents, ranging from minute openings to some of several feet in diameter at the mouth appeared to the west and north. The steamy condition of the ground at this time is shown in Figure 10, and the escape of steam under heavy pressure from outlet pipes of the drilling rig is shown in Figures 11 and 12.

It was now found that some solid obstacle blocked the hole at a depth of 36 feet, and repeated attempts to get past it were unsuccessful. It appeared to be an end of broken drill casing. The uprushing

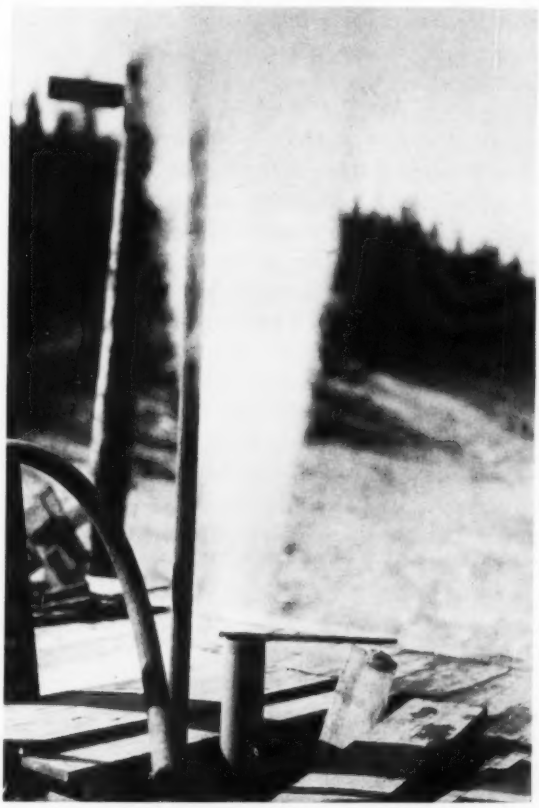

FIG. 12.—Close-up view of high pressure steam escaping, Norris hole

steam had carried with it a considerable amount of grit and sand, and our supposition was that this had acted as a sand blast on the casing, wearing it so thin that it was not able to stand the pressure when the valves were closed. It had burst, and the sections above and below the break were no longer in line. Various attempts were

made for several days to remedy the situation, but nothing gave hope of success, and it was finally decided that it was futile to continue.

These various difficulties illustrate factors that will have to be taken into account in any attempt to utilize underground sources of power under similar conditions.

We felt some apprehension that when the valves were finally closed and the orifice was sealed there would be an upheaval. As a possible safeguard against this, five tons of cement batter were rapidly pumped down the hole. Whether this was helpful or not, there were no serious results. Two or three hours after the valves were closed a large blowout appeared about 75 feet away, and the next day there was another at a distance of 50 feet. These vents continued active, but the discharge was not violent, and they did not increase in size. The drill hole was sealed at the surface with concrete, and there has been no further outbreak.

The material encountered in the uppermost part of the drill hole appeared to be rock débris that had been washed over the site. This was succeeded by a mixture of boulders and sand that probably represented disintegration of rock in place, and finally solid rock was entered. Naturally there was no sharp boundary separating one class of material from another. A depth of 58 feet was reached before the rock became continuously solid. Above this, an occasional boulder had been passed through, only to be succeeded by disintegrated material.

The odor of H_2S was more prominent during the drilling of this hole than at the one in the Upper Basin, but was not strong. There was rather more secondary pyrite in the rock here, and in some of the cores numerous grains of it may easily be seen with a hand lens. In spite of this, many of the cores from all depths show irregular stains or films of ferric oxide.

The presence of the ferric oxide, though a minor matter, has been a puzzle. All the evidence is against the idea that the solutions at these depths are oxidizing, and if it had been formed at a previous time (possibly by weathering of the rhyolites shortly after they were laid down) it does not seem that it should have persisted, especially in the presence of solutions that deposited pyrite.

In some instances a part, at least, appears to be the result of the

grinding action of the drill bit or core barrel against the core, as the outside of cores is more deeply stained. This may account for all of it, though one would hardly have supposed that the pressure of the feed water would force it into the interior of cores in the manner in which it occurs;[17] but the wear on the bit was rapid, the particles of iron or iron alloy of this origin were very fine, and the cores are porous. The oxide stains are always in amorphous films, never crystallized.

TEMPERATURES AND PRESSURES

The great difficulties encountered in drilling, arising from the fractured or disintegrated condition of the rock mass and the high pressures and temperatures of steam, had their counterpart in obtaining measurements of temperature and pressure. Small movements of rock adjacent to the drill hole occurred a number of times. The projections thus formed were easily reamed out by the drilling tools but they held up the thermometer, or caused difficulty in getting it back after it had been lowered. For a period, the withdrawal of the drilling tools was followed by a dropping or washing in of disintegrated rock or sand, which filled the lowermost 40–80 feet of hole, and prevented the thermometer from reaching the bottom. This condition was partly remedied by driving the outer casing deeper, as most of the trouble originated at a weak place about 60 feet down.

After high pressures had been struck, it was necessary to place the thermometer in a special receptacle at the end of the drill rods, and lower the whole string and raise it again for a temperature measurement. This required that the line of rods should be assembled in lowering, and disassembled in raising for each measurement. With high pressure steam rushing out, this was a slow and difficult operation for even experienced drill men. Several hours were necessary to complete an operation of this kind.

The gate valve that was first installed at the surface for shutting off steam was one that was rated at 125 pounds maximum operating pressure. Our experience in drilling the hole in Upper Geyser Basin had led us to suppose that this would be sufficient. When higher

[17] The feed water for drilling was pumped from Nuphar Lake, almost at the place where drainage from the acid area around the drill site enters. The water was probably slightly acid and slightly oxidizing.

pressures were suddenly encountered, a gate valve of greater capacity was sent for and installed. In the meantime it was not considered advisable to close the valve completely; therefore no measurements of full pressure could be taken.

When the pressures became extreme, it seemed likely that shutting off the steam completely for more than a short time would result disastrously. During most of the time, the steam was allowed to escape without check, and it was only shut off during the interval required for making a measurement of pressure. Under these circumstances it is doubtful what the steam pressure that was measured really represented. The steam was brought into the hole by fissures from some deeper source, which had been partly depleted by long-continued outflow. Temperatures were doubtless affected by the expansion of the steam during its escape. The maximum temperature obtained was 205°, on September 16 at a depth of 246½ feet. This can hardly represent the temperature of steam at the gauge pressure of 297½ pounds that was obtained a few hours earlier. The measured pressure is greater than that of saturated water vapor at this temperature. That it should actually be greater is impossible. Under the conditions described, it is easy to see how discordances between pressure and temperature might arise.

Because of these circumstances, the following tabulation of temperatures and pressures is supplemented with descriptive comments in order that the conditions under which they were taken may be understood. The figures given for pressure are the readings of the gauge. Approximately 15 pounds per square inch should be added to obtain absolute pressures.

August 12: Depth = 20 feet. Temperature = 95°.

August 19: Depth of hole = 71 feet, but thermometer was held up by a projection at 56 feet. Temperature = 115°.

August 23: Depth of hole = 129 feet, but with filling of sand for 40 feet. Pressure = 15 pounds. Thermometer at 72 feet = 121°.

August 23–31: Measurements impracticable because of sand washing in when drill tools were withdrawn.

September 1: Depth = 202 feet 9 inches. Pressure = 46 pounds.

September 3: Depth = 215 feet. Thermometer at 200 feet = 141°.

September 4: Depth = 220½ feet. Temperature = 135°. Pressure = 35 pounds.

September 5: Depth = 230 feet. Thermometer was stopped at 107 feet. Tem-

perature there = 144°. Pressure = 55 pounds. At about 220 feet a fissure had been passed through, and the feed water was "lost" for 3 hours.

September 6: Depth = 242 feet. Pressure = 48 pounds. A few hours later pressure = 112 pounds.

September 7: Depth = 243 feet 7 inches. Pressure = 130 pounds with outlet valve not entirely closed. (Low-capacity valve was in use.)

September 8: Depth = 243 feet 7 inches. Thermometer was lowered and became firmly caught. It was fished out on September 10, and read 146°, but had probably been jarred.

September 9: Depth = 243 feet 7 inches. Pressure = 181 pounds. A later reading gave 170 pounds.

September 10: Depth = 243 feet 7 inches. (No drilling was being done. Conditions were being made more secure.) Pressure = 93 pounds at 10:07 A.M., 88 pounds at 10:27 A.M. Temperature = 159°.

September 11: Depth = 244 feet 10 inches. Pressure = 89 pounds. At a depth of 246 feet 6 inches a pressure of 225 pounds was suddenly struck. Valve closed for only a moment in taking pressure.

September 12: Depth = 246 feet 6 inches. In the morning, after valves had been half-closed since the day before, pressure = 225 pounds. At 2:00 P.M. the pressure gauge was attached, and outlet valves were closed for a moment. Indicator rose at once to 250 pounds. (See Figs. 11 and 12, representing outrush of steam at the 3-inch pipe at this time.)

September 16: Depth = 246 feet 6 inches. A block of concrete was being laid at the surface. Since September 11 the valves had been wide open most of the time. At 9:35 A.M. the gauge was attached. The indicator rose almost at once to 297½ pounds. At 9:40 A.M. pressure = 276 pounds, at 9:44 A.M. 268 pounds, at 9:50 A.M. 265 pounds. In the afternoon the thermometer was lowered on drill tools to the bottom of the hole, was left for 10 minutes, and then pulled up. Steam valve at the side was wide open during the operation. Temperature registered 205°.

September 17: Depth = 246 feet 6 inches. Pressure at 9:17 A.M. = 250 pounds. Outlet valve had been wide open all night, but was nearly shut off at 8:20 A.M. For taking pressure it was shut off completely.

September 18: Depth = 246 feet 6 inches. Attempts to drill had been unsuccessful. A pressure reading this morning after full escape of steam during night gave 250 pounds. Later, 2 feet was drilled. A pressure reading was taken after steam had blown off all night and all day, except that it was shut off for 40 minutes before reading was made. This gave 277½ pounds, and remained constant for 30 minutes.

September 20: At 8:00 A.M., depth = 248 feet 11 inches. Pressure = 250 pounds. At 3:00 P.M., depth = 254 feet 10 inches. After valves had been closed for 1 hour, pressure = 265 pounds.

September 22: Depth = 263 feet 4 inches. Pressure = 265 pounds.

September 23: Depth = 264 feet 10 inches. Pressure = 252½ pounds.

September 24: Pressure dropped to 145 pounds, then to 125, then to 113. Blowouts appeared in the surrounding ground. Casing pipe ruptured 36 feet below surface.

October 2: Attempts to continue work were abandoned.

This somewhat unsatisfactory record of temperatures is represented in Figure 13 by the line *AA*. The depths and temperatures to which *AA* is referred are shown by the scales at the left and top, respectively. The line *BB* shows the pressures of saturated water vapor

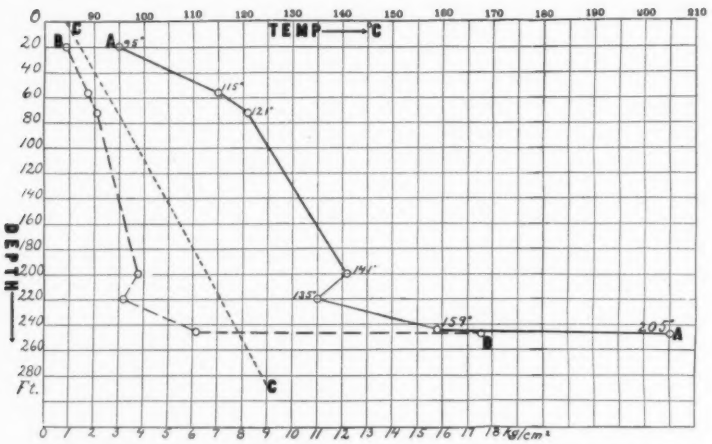

FIG. 13.—Graphs of temperatures and pressures in drill hole of Norris Basin. *AA* = measured temperatures, *BB* = vapor pressures of saturated water vapor at the measured temperatures, *CC* = hydrostatic pressures of a water column at the various depths.

at the measured temperatures, the scale of pressures for this graph being given at the bottom of the diagram. *CC* represents the calculated pressures of a column of water at various depths from the surface downward, allowance being made for the decrease of density with increasing temperature (as measured), and the weight of the atmosphere having been added.

ALTERATION OF FELDSPAR TO CLAY IN NORRIS CORES

At the surface in Norris the phenocrysts of orthoclase have been completely leached out of the rhyolite, leaving little cavities of char-

acteristic shape, and the whole rock looks leached. The analysis of this rock is given under Y.P. 654 in Table XIII. The analysis shows a material containing little but SiO_2, Al_2O_3, and H_2O. In the thin section quartz phenocrysts are numerous, but those of feldspar have disappeared. In the groundmass small aggregates of secondary tridymite are visible, but most of the groundmass is so fine grained that it appears nearly isotropic, and optical identification of minerals is impossible. When a little of the powder is tested in a solution of malachite green in glycerine the particles are stained, indicating the presence of clay.

This thorough leaching is the effect that we should expect from the attack of the acid waters that are present at the surface here. According to Allen's explanation as to the manner in which H_2SO_4 has been formed, the acidity should disappear at a moderate depth, and the waters below should be alkaline.

In the core from a depth of 28 feet only a few feldspar phenocrysts have been removed, and at 34 feet they are nearly intact. This rapid decrease of alteration seems at once to confirm Allen's view, but when the cores from greater depths are examined a few feldspar phenocrysts in all are found to have been replaced by a clayey substance, and this continues without diminution to the bottom of the hole. When this was first observed it seemed to be contrary to what we should expect. It will be recalled, however, that in order to account for bicarbonates of the alkalis in the thermal waters, Allen[18] has reasoned that there must have been an attack by CO_2 upon the wall rocks at depth. This action also would presumably result in the formation of clay. In the cores from the Upper Geyser Basin no evidence of this action was found, and it was necessary to suppose that it took place at a greater depth than was reached by the bore hole. In the Norris cores it seems possible that we have evidence of this process. We should then have formation of clay by two different and unrelated processes: near the surface by the attack of acid sulfate, and at depth by the attack of CO_2 in an alkaline solution. If means could be found to confirm this, we should have corroborative evidence of the correctness of Allen's inductive reasoning. Confirma-

[18] E. T. Allen, "Geyser Basins and Igneous Emanations," *Econ. Geol.*, Vol. XXX (1935), p. 11.

tion would also be of interest in throwing light on the more general problem of the character of solutions in which clays have been formed. A full knowledge of this would have important applications.

Within the last few years valuable contributions have been made by Ross, Kerr, Larsen, Wherry, and Shannon in determining the constitution of clays and in establishing diagnostic properties. It is still almost impossible, however, to discriminate among the minerals by optical methods when they are in their usual fine-grained condition, and the only practicable means is by their X-ray patterns. For the study of the Yellowstone material, it was through the kind assistance of E. Posnjak, who had previously made many X-ray studies of clays, that results of value were obtained. The optical data secured by the writer are in accord with Posnjak's determinations, but by themselves would not have been sufficient.

There is considerable evidence in descriptions given in the literature that kaolinite is formed in acid solutions. The beidellite-montmorillonite-nontronite group seems to be indicative of alkaline conditions, but the evidence is less positive.

Allen has collected several samples of clay in association with thermal springs at the surface in Yellowstone, and these have been examined optically by Merwin and by X-rays by Posnjak. Kaolinite has been found to be associated with acid waters and beidellite with alkaline.

Four samples of material from the Norris hole were examined for the writer by Posnjak.

Y.P. 654. Thoroughly leached surface rock. As the feldspar casts were empty, material from the fine-grained groundmass was used. This would be expected to contain, in addition to clay, primary and secondary silica minerals. Posnjak found kaolinite, together with much cristobalite, quartz, and tridymite. (Note: Cristobalite has not been recognized microscopically in any of the Norris cores. This surface specimen, however, is different from any other. Possibly cristobalite was formed by the decomposition of beidellite to kaolinite.)

Y.P. 680. From depth of 95 feet. White clay replacing phenocrysts of feldspar. It was found by X-rays to be mostly kaolinite, with a little residual feldspar, but it may contain some beidellite also.

Y.P. 735. From 198 feet. White clay from feldspar casts. It con-

sists essentially of the beidellite type of clay. Grains of feldspar are present.

Y.P. 762. From 243 feet. Cream colored clay from feldspar casts. It is the beidellite type of clay. Probably some feldspar is present.

These results, considered in connection with the character and occurrence of the materials examined, show that in the surface rock feldspars have been wholly kaolinized. In the core from 95 feet only a few feldspar phenocrysts and (as we shall see) a little of the groundmass have been altered, and the clay is probably a mixture of kaolinite and beidellite. At greater depth the alteration has produced beidellite only.

In making the examinations Dr. Posnjak was unaware as to the kind of clay that was expected in the specimens. It was gratifying that his results were in harmony with what had been deduced from the conception of processes derived from other evidence. Only a minor discrepancy was found in that the kaolinite at 95 feet indicates that acid solutions penetrate somewhat deeper than we had supposed.

The effect of acidity is thus apparent in the formation of kaolinite as far down as 95 feet, and probably extends somewhat beyond that, but only a small amount of feldspar has been altered at this depth, and therefore the action is not vigorous. This does not imply, of course, that meteoric waters do not descend below the shallow zone of acid leaching. They doubtless descend much deeper, but at such a slow rate that the acidity that they tend to impart is overcome by the alkalinity of the bicarbonates. Probably the body of underground water is more nearly stagnant at Norris than in the Upper Basin; therefore, it does not carry away so rapidly the heat imparted to it from below, and its ultimate depth is less. This, in turn, brings the zone of CO_2 leaching within the levels reached by the drill hole, whereas in the Upper Basin it lies below the depths explored.

The analysis of the surface rock (Y.P. 654 in Table XIII) shows that the strong bases have been almost completely removed. Four analyses of cores from various depths are given in Table XIII. Their molar ratios of strong bases to alumina are indicated in Table X.

These ratios are not so close to unity as in the cores from the Upper Basin. Inasmuch as the rocks at Norris were of more strictly

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The analysis of the surface rock (Y.P. 654 in Table XIII) shows that the strong bases have been almost completely removed. Four analyses of cores from various depths are given in Table XIII. Their molar ratios of strong bases to alumina are indicated in Table X.

These ratios are not so close to unity as in the cores from the Upper Basin. Inasmuch as the rocks at Norris were of more strictly

rhyolitic character, the original ratios should have been closer to unity, and if the only reaction had been substitution of potash for soda and lime, as in the Upper Basin, the 1:1 ratio should have been maintained. Superimposed on this process there has been a later attack—by sulfuric acid near the surface and by carbonated waters at greater depth.

In Table X, it is to be observed that the core from 70 feet has suffered the greatest loss of bases (next after the surface rock). The

TABLE X

MOLAR RATIOS OF $\frac{K_2O+Na_2O+CaO}{Al_2O_3}$ IN NORRIS CORES

Specimen	Depth (Feet)	Ratio
Y.P. 669	70	0.860
Y.P. 709	157	.904
Y.P. 750	221	.933
Y.P. 776	264	0.908

analysis confirms the X-ray finding that acid leaching reaches to this depth, though quantitatively its effects are small.

In the three analyses from 157 feet to 264 feet no trend toward greater or less leaching appears. These are apparently beyond the influence of acid leaching, and accordingly the clay at these depths is found to be of the beidellite type.

From a number of specimens in which clay appears (from 221 to 236 feet), a little was obtained for analysis by working it out of the casts with a fine needle. The amount was only 0.0925 gram, and it was not possible to obtain it wholly free from small grains of quartz and feldspar. The analysis should therefore be regarded as only an approximation to the composition of the clay. The results were as shown in Table XI.

Alkalis were not determined, but the amount must be small. The molar ratio of silica to alumina is high (0.930 SiO_2 to 0.226 Al_2O_3 or 0.259 $Al_2O_3+Fe_2O_3$), and this is in accord with beidellite (3:1 molar) rather than kaolinite (2:1). Doubtless part of the silica is due to the inclusion of grains of silica minerals and feldspar, but it is not likely

that so much is present as to give the great excess over the kaolinite ratio. Ferric oxide, lime, and magnesia are present in considerably greater percentages than in the rocks as a whole, and this again is in accord with the characteristics of the variable montmorillonite-beidellite-nontronite series.

In the examination of cores with the binocular magnifier most of the beidellite that is discernible occupies casts of feldspar phenocrysts, but in a few specimens it may be seen as a coating over the

TABLE XI

PARTIAL ANALYSIS OF CLAY REPLACING ORTHOCLASE PHENOCRYSTS

	Per Cent
SiO_2	55.78
Al_2O_3 (+P_2O_5+TiO_2)	23.03
Fe_2O_3 (total iron)	5.05
CaO	1.08
MgO	1.13
H_2O-	5.30
H_2O+ (by ignition)	5.08
Total	96.45

little nests or vugs of secondary minerals. In thin sections it is observed to replace feldspar phenocrysts wholly or partially, and may be recognized occasionally in the groundmass. Probably much more is present than is distinguishable.

The optical properties of the clays in the specimens identified by Dr. Posnjak, and in a number of others, were determined as far as possible, but the minuteness of the crystalline fibers made it difficult to obtain satisfactory results. Moreover, the montmorillonite-beidellite-nontronite series is of variable composition, and though the members are united by a certain type of atomic structure and by certain chemical characteristics, the indexes of refraction vary widely, and extend above and below those of kaolinite. In these rocks the beidellite clays in the casts are usually white, but some are light brown or light green. From appearance and optical measurements they probably vary from nearly pure hydrated aluminum silicate to those members in which considerable ferric oxide and magnesia are present. Nothing inconsistent with Dr. Posnjak's determinations

was found, but, from the nature of the case, not more than mild support could be expected.

Destruction of feldspars and formation of the beidellite type of clay is apparently now the active process in the deeper levels at Norris. It is going on slowly and has not yet brought about sufficient changes to obscure previous processes, but the clay penetrates almost everywhere, and it is revealed by staining in almost every specimen when the powdered rock is placed in a solution of malachite green. The positive staining tests in the Norris cores are in contrast with the negative tests in the cores from the Upper Basin. In the latter cores no staining that appeared to be attributable to clay minerals was observed, though many tests were made. This confirms the petrographic and chemical evidence that little or no clay is present in those rocks. It appears probable, therefore, that in the Upper Basin the reactions by which the carbonated waters have acquired their complement of alkalis have always taken place below the zone explored by the drill hole.[19]

Presumably the destruction of feldspars to form clay is incompatible with the main process that was previously active at Norris and is still going on in the Upper Basin, which results in the formation of secondary orthoclase.

An interesting feature is the close association of beidellite with pyrite. In innumerable instances the casts of feldspars replaced by clay contain bright little cubes of pyrite, and portions of groundmass that contain an unusual amount of clay are likely to have much pyrite. This does not always hold, but the association is so common that one feels that the same alkaline solutions that formed the clay introduced the pyrite. Several cores are cut by narrow bands of clay in which slickensides indicate movement, and these clays are quite

[19] In his paper "Geyser Basins and Igneous Emanations" (p. 11) Allen wrote: "In the alkaline waters of the Norris Basin, the bicarbonates are much lower in concentration (averaging only one-eleventh that in the Upper Basin) than they are in any of the other geyser basins, and the free gas is greater in amount. Perhaps this means that less carbon dioxide is absorbed in reaction with the rock because its path through ground water is shorter."

With the identification of beidellite in the Norris cores, it appears that the zone of attack by CO_2, for which Allen had deduced a shallow depth, actually comes within the region of observation.

highly pyritiferous. They resemble the mineralized gouges of ore veins. In two or three very narrow veins observed in thin sections, pyritiferous clay is present among broken and slightly displaced phenocrysts of feldspar.

Many phenocrysts of feldspar that still appear glassy have been attacked along cleavage planes so that they are ready to fall apart with small pressure. Others that appear intact have had films of pyrite introduced.

In the Upper Basin, the thermal waters, according to Allen's analyses, show an average molar ratio of 34 Na_2O to 1 K_2O. This gives little or no clue to the relative concentration of the two alkalis in the halide emanations from the magma, for the proportions have been greatly changed by subsequent reactions. After the emanations are absorbed by ground water their first reaction is attack upon the wall rock (because of their content of CO_2). Since this action is very little selective, the two alkalis are taken up in about the proportion in which they are present in the feldspars.[20] Most of the free CO_2 is thus used up and so much alkali is acquired that the molar ratio of HCO_3[21] to Cl in the waters of the Upper Basin becomes nearly 1:3 (according to Allen).

At a higher level the waters effect a base exchange with the feldspars of the rock, and in this process they lose potash and gain soda and a little lime. When they issue at the surface, after having undergone these modifications, the K_2O/Na_2O ratio has been altered to such a degree that it now has no relation to the original ratio.

In Norris Basin the reactions going on at present are much simpler, and it seems possible to derive conclusions as to the character of primary emanations. There is an element of uncertainty in any case of this kind as to whether we have taken account of all the elements of the problem, but the data we have are so much more complete than is usual in such cases, and have led to such consistent conclusions that we feel justified in pursuing the inferences further.

[20] We should recognize the possibility that at the high temperatures at which carbonate attack is taking place, below the zone of exploration in the Upper Basin, muscovite or sericite and not clay may be the residual mineral formed. This will not greatly affect the point that is to be brought out.

[21] HCO_3 expresses the bicarbonate radical to which alkalis are attached. It does not include free CO_2.

In Norris Basin the attack of CO_2 upon the wall rock is much less than in Upper Basin, and most of the CO_2 present is uncombined. Allen finds that in the alkaline waters of Norris[22] the average of seven analyses gives a molar ratio of HCO_3/Cl of only 0.028:1 (with limits of 0.0136:1 to 0.044:1). The bicarbonates are much lower in concentration in these waters than they are in any of the other geyser basins, and the free gas is greater.

We shall assume that the small proportion of alkalis present as bicarbonates in the Norris waters was acquired from the wall rock and that the alkalis present as chlorides were derived from the magma. If we subtract bicarbonate alkalis from total alkalis we shall then have the potassium and sodium chlorides in the proportion in which they are given off by the magma.

To the HCO_3 radical we allot K and Na in the proportions in which they exist in Norris rocks analyzed by the writer, and find that in 100 molar parts of alkalis present in the waters, there are $1.25NaHCO_3$ and $1.45KHCO_3$.

We have the further fact (from Allen) that the total Na:total K in alkaline waters of Norris is about 90:10. We subtract from these respective figures 1.25Na and 1.45K present as bicarbonate, and obtain 88.75Na and 8.55K that emanate as chloride from the magma, or 91.2 (molar) per cent NaCl and 8.8 per cent KCl.

This high sodium content of the magmatic emanations is in accord with common observation regarding the composition of volcanic sublimates. If, however, the conclusions are applied to the Upper Basin, it may seem surprising that the evolution of such vapors should set up a chain of reactions that finally results in the replacement of sodium by potassium in the rocks. We should remember, however, that the metamorphism found in the bore hole in the Upper Basin is the end-result of a series of processes. We should supplement the information obtained there with what we find at Norris. Then, if our deductions are correct, we perceive that the potassium that is so important in substitution processes in the feldspars is not derived (except, per-

[22] Both acid and alkaline springs occur at Norris. The acid springs acquire their character by the superficial processes that have been described. The alkaline springs are believed to represent deep-level waters that reach the surface without having been much modified by acid reactions.

haps, in small degree) from the igneous body but from carbonate attack upon feldspars of roof rock at a lower level than the bore hole attained. In short, the potassium is picked up at lower levels by one process and is deposited at higher levels by another. The surprising thing (not dependent upon theory or inference) is that in the upper levels substitution of potassium for sodium goes on until the waters are so depleted of potassium and enriched in sodium that the ratio of the two becomes 1:34. This, however, is established with almost as much certainty as if the experiment were performed in the laboratory.

GENERAL CHARACTERISTICS OF NORRIS CORES

The formation of clay from feldspars is apparently the process now going on at Norris, but it has not yet advanced so far as to destroy or greatly obscure the effects of previous processes. Through careful study these have been found to be very similar to the processes of base exchange in feldspars in the cores of the Upper Basin, but significant clues are less in evidence at Norris, and if study of these cores had been attempted before the knowledge gained from the cores of the Upper Basin had been acquired, it is doubtful if their story could have been deciphered.

In the Upper Basin the original phenocrysts were seen to be incongruous with the present composition of the groundmass, as determined by analyses. It was fortunate, also, that a specimen of wholly unaltered core was preserved for comparison with altered cores. The retention of perlitic structures of the original glasses was also of much assistance. A vital feature was the maintenance of the 1:1 ratio of strong bases to alumina. In the Norris cores guides of this kind are almost lacking.

The original rocks were sodic rhyolites. They carried plentiful phenocrysts of quartz and orthoclase in a groundmass that is usually distinctly banded or laminated (Fig. 14).

The lamination may indicate flowage of liquid lava, but, in many specimens, features are visible in thin sections that strongly suggest welded tuffs.[23] A few cores that have an almost uniform groundmass may have been vitrophyres.

[23] J. P. Iddings, in his description of the rhyolites of Yellowstone Park (*op. cit.*, pp. 356–432), mentions welded fragmental lavas in several places. The present writer has

In rocks of this rhyolitic composition, replacement of soda and lime by potash in the groundmass feldspars, and addition of silica, would not have given a result recognizably discordant with the phenocrysts of quartz and orthoclase. As if to make interpretation more difficult, the secondary orthoclase of the groundmass is not a pure potash orthoclase, as it is in the Upper Basin, but is a variety

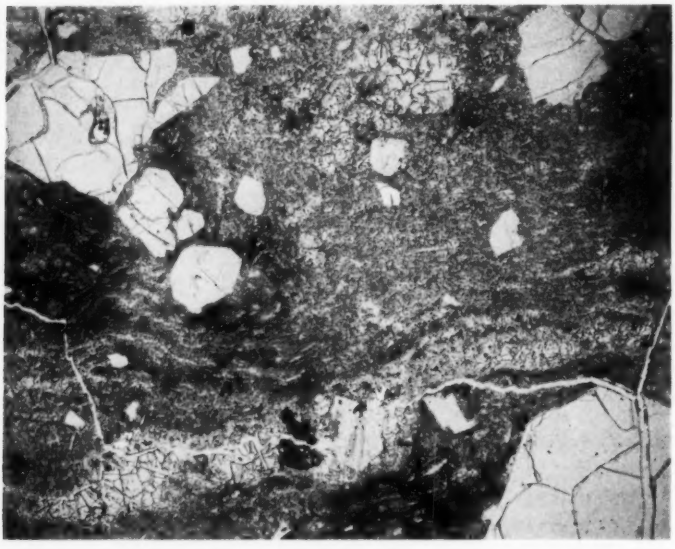

FIG. 14.—Photomicrograph of lenticles and pockets of secondary tridymite and orthoclase replacing laminated groundmass of rhyolite. Y.P. 733, from a depth of 196 feet in Norris Basin. ×16 diam. Ordinary light.

that contains a large amount of soda, as was shown in an early section dealing with the properties of orthoclase. From the evidence given in the section just preceding this one, it seems likely that at the period when this metamorphism was going on at Norris the solutions coming up from below had not acquired so much potassium *bicar-*

seen other examples, and they seem to be of rather common occurrence. In hand specimens many of them appear hard and dense, and their fragmental character would hardly be suspected.

bonate here as in the Upper Basin. Whether this was the important factor in producing a more sodic orthoclase cannot be stated.

In order to obtain certain essential information as to the completeness of metamorphic replacement in Norris cores, which could not be determined by optical examination of the fine-grained groundmass, a typical specimen, Y.P. 750, from a depth of 221 feet, was selected for special study.

Part of the groundmass of this core is seen to be dark gray, lithoidal, with no appearance of alteration, even when it is closely inspected with a binocular magnifier. Another part is white, with minute pores and small cavities lined with little crystals. In places the two facies are intimately mixed. If alteration has occurred, it is evidently the white portion that has been changed. Powdered material from this, examined microscopically, shows orthoclase, tridymite, and quartz. These minerals are not obviously discordant with the original rhyolitic composition, and their slightly coarser development might indicate merely recrystallization without change of composition. An analysis of the core material, comprising both facies in their natural intermixture (see Y.P. 750 in Table XIII), shows high potash, but the composition would not be commonly regarded as abnormal for a rhyolite.

Some of the core was coarsely crushed, and fragments of the unaltered original phenocrysts (quartz and orthoclase) were picked out. Several hundred fragments were required to give enough for analysis. The results (Y.P. 750A in Table XIII of analyses) have been cited on an earlier page to illustrate the relation of refractive indexes of orthoclase to potash-soda content. Their citation at this point is to serve, in connection with other analyses, the more important purpose of showing distribution of soda and potash between phenocrysts and groundmass. The analysis gave 4.61 per cent K_2O and 2.72 per cent Na_2O, or a molar ratio of K_2O/Na_2O equal to 1.140.

A second analysis (Y.P. 750B), determining alkalis alone, was made of the porous, white, finely crystalline facies of the groundmass. In this, $K_2O = 8.33$ per cent, $Na_2O = 3.58$ per cent, molar ratio $K_2O/Na_2O = 1.517$. The ratio of potash to soda is decidedly higher in the altered groundmass than in the phenocrysts.

The significance of these relations is this: When phenocrysts of

orthoclase crystallize from a potassic-sodic magma, they may carry much soda, but the ratio of potash to soda should be higher in the phenocrysts than in the groundmass, and presumably that was the relation in the original rock, but now the groundmass shows a higher K_2O/Na_2O ratio than the phenocrysts. Evidently the groundmass has received accessions of potash.

This conclusion is confirmed when we compare these figures with the molar ratio in the whole rock—1.229. By a simple calculation we find that if the unaltered portion of the groundmass is assumed for the moment to have the same ratio of K_2O to Na_2O as the phenocrysts, it would require 77 per cent of unaltered groundmass plus phenocrysts, mixed with only 23 per cent of altered groundmass, to give the ratio that was found for the whole rock. Inspection of the specimen shows that the amount of altered groundmass is much greater than this; therefore unaltered groundmass plus phenocrysts must have a lower ratio than was assumed, which in turn means that unaltered groundmass has a lower ratio than phenocrysts—the relation appropriate to the original rock.

We have, therefore, a rock in which parts of the groundmass (the gray, lithoidal portions) have been changed little or not at all, whereas, in other parts (white, finely crystalline), potash has replaced much of the soda of the groundmass.

Complete analyses of other typical specimens of Norris cores are shown in Table XIII under Y.P. 669, Y.P. 709, and Y.P. 776. All have a close similarity to Y.P. 750. The last has a somewhat higher ratio of potash to soda than the others, and was therefore the most suitable for demonstrating results of alteration, but the effects do not vary much. Practically all cores from this hole show irregular intermingling of altered and unaltered portions. In all respects, except the acid leaching of feldspar near the surface, there is little difference among them from top to bottom.

The fact that exchange of alkalis in the feldspars of the Norris cores has been only partially effected illustrates the slowness with which metamorphic processes have operated. A period of action of probably many thousands of years was not sufficient to bring about equilibrium.

It is believed that in the Norris rocks penetrated by the drill hole

the substitution of potash for soda and lime in the feldspars has now ceased, but the results of this process are still the most prominent effect. They are similar to those described in the cores from the Upper Basin, but differ in details.

The principal difference is that the newly formed orthoclase is a highly sodic variety, as already described. There has been much deposition of silica here, as in the Upper Basin, but tridymite is the chief silica mineral. There is much quartz also, and a little opal is found at various levels, apparently of late deposition.

Calcite, found in many cores of the Upper Basin, has not been detected at Norris. In the analyses (Table XIII) CO_2 is indicated as present to the amount of 0.04 per cent in Y.P. 750 and 0.01 per cent in Y.P. 776, but these amounts are within the limits of uncertainty of the analytical method, and its presence is doubtful. We should not expect calcite to be deposited by waters in which the amount of free CO_2 is large. Calcium bicarbonate would remain in solution.

Zeolites are absent or are doubtfully present in small quantity. The same is true of chlorite.

Pyrite is not abundant, but is found in many cores. There is much more than in the Upper Basin. The 0.41 per cent sulfur in the analysis of Y.P. 750 shows rather more than the average amount of pyrite. It occurs as small, bright cubes or as films. Though plainly secondary, and associated so frequently with clay minerals as to indicate introduction by the same waters that formed the clay, it is frequently seen along cleavage planes within glassy phenocrysts of feldspar.

By far the most abundant secondary minerals are tridymite, quartz, and sodic orthoclase (Fig. 15). These occur as metasomatic replacements of the groundmass and, in somewhat larger crystals, as the lining of vugs. Many of the cores have these vuggy cavities, which are usually in the form of small lenticles, roughly parallel with the primary lamination. Their length is a few centimeters or less. Some of the cavities are simply irregular holes. All are plainly due to solution. Clean-cut veins of secondary minerals, such as are a common feature in the Upper Basin, are absent here.

The cavities are lined with secondary quartz, tridymite, and orthoclase. Quartz is in stout crystals, combining prism and pyramid.

Tridymite and orthoclase are in thin, tabular crystals, a few tenths of a millimeter in greatest diameter. The two look so much alike that they are indistinguishable megascopically, but they appear very different under the microscope. Their optical identification has been confirmed by X-ray examination by E. Posnjak.

In polarized light, the tabular faces of the orthoclase are found to be nearly perpendicular to an obtuse bisectrix or an optic normal,

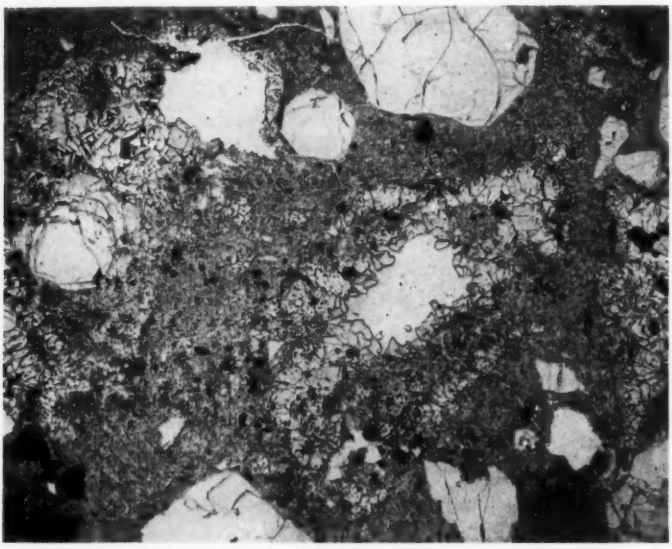

FIG. 15.—Photomicrograph of vugs and pockets of secondary tridymite and orthoclase, in relatively large crystals, replacing fine-grained groundmass of rhyolite. The phenocrysts are quartz. Y.P. 695, from a depth of 128 feet in Norris Basin. ×16 diam. Ordinary light.

hence these faces are inferred to be [010] or [001]. All the secondary orthoclase at Norris is of this tabular form, whereas in the Upper Basin most of the crystals are nearly equidimensional or slightly elongated along the vertical axis.

In addition to these crystal-lined cavities, spongy aggregates or irregular pockets of the same minerals in smaller crystals occur. Seen in thin sections they may be quite sharply defined or may merge

with the finer-grained groundmass. No definite order of succession of the three minerals is apparent.

Chemically, these rocks were originally made up chiefly of free silica and potassium-sodium aluminosilicate.

These compounds may have been present in the groundmass as crystallized minerals or they may have been occult in a glass. In either case the aluminosilicates responded to the action of thermal waters of a composition that required for equilibrium a more potassic orthoclase than that originally present.

The formation of tridymite is not so simply explained. If the free silica of the primary groundmass was present as quartz, this stable form would certainly not yield to the unstable form tridymite. There are two ways in which the deposition of tridymite might be accounted for. First, the free silica may have been dispersed in a glass, in which case crystallization resulted in the formation of a silica mineral that, while it was not the most stable phase, was more stable than the previous one. Second, quartz may have been originally present in the groundmass, but it may have been picked up by solutions at one place, carried onward, and deposited under slightly different conditions as tridymite at another place, especially in cavities that had been formed by solution. It is plain that both tridymite and secondary orthoclase did migrate into these cavities, and many cavities are now lined or filled with these minerals. Even original phenocrysts of orthoclase are cut by microscopic veinlets of tridymite and orthoclase, and the irregularity and lack of match of the two walls indicate that the space was originally formed by solution.

Thus it appears that whether the original silica was present in glass or as quartz it moved from one place to another, and was frequently accompanied in its migration by orthoclase. The high values of silica indicate, too, that some was brought up from below.

Many phenocrysts of quartz exhibit a rather curious effect of another sort. In the thin sections they are seen to be replaced at the borders and even in their interiors by little crystals of secondary orthoclase. These are commonly lathlike forms (cross sections of thin plates), and are sharply bounded by crystal faces. They were evidently not deposited in cavities, for no space is left unoccupied by each euhedral little crystal. What caused the stable quartz to yield

to another stable mineral, orthoclase? The explanation seems to lie in what may be called a *second order* of instability of the quartz at the places where pressure was exerted (and solubility increased) by the crystallizing force of the growing orthoclase. Some of these little crystals lie isolated within phenocrysts of quartz, and the channels by which the depositing solutions gained access are imperceptible.

Phenomena of this kind imply processes that are sometimes neglected in the theoretical consideration of metamorphism, as being of too small an order to be taken into account. Here quartz, which is shown by its general relations to have been a stable mineral, has given way before the advance of orthoclase; and the solutions that deposited the orthoclase in fairly large crystals entered the quartz phenocrysts through submicroscopic cracks. Such processes are probably very common. They add to the difficulty of interpreting the results of metamorphism, but they have to be recognized.

In order to establish that zeolites, chlorites, alunite, and calcite are not present in the fine-grained groundmass of the cores, a portion of the powder of Y.P. 750 was extracted with boiling dilute HCl, followed by Na_2CO_3, in the manner described for Y.P. 273B from the Upper Basin (see Table IV).

TABLE XII

EXTRACTION ANALYSIS OF Y.P. 750C

Constituent	Per Cent	Constituent	Per Cent
SiO_2	1.35	MgO	0.01
Al_2O_3+ P_2O_5+TiO_2	0.24	Na_2O	.08
		K_2O	0.14
Fe_2O_3 (total iron)	0.37	Total extracted	2.21
CaO	0.02		

The very small amount of soluble material extracted by this treatment shows quite definitely that there is present no significant quantity of certain minerals that might be considered likely to occur and that might have escaped detection in the microscopic examination.

ANALYTICAL RESULTS

In Table XIII the analyses of Norris material are assembled. Reference has been made to these at various places in the text. Little

further comment is required, except to indicate what material they represent. Y.P. 654 is surface rock leached by acid sulfate waters. Y.P. 669, Y.P. 709, Y.P. 750, and Y.P. 776 are cores from various depths, selected as typical of core material. There is little difference among them. All are rhyolites, in which part of the soda and lime of

TABLE XIII

COMPLETE AND PARTIAL ANALYSES OF NORRIS DRILL CORES

	Y.P. 654 (Surface)	Y.P. 669 (70 ft.)	Y.P. 709 (157 ft.)	Y.P. 750 (221 ft.)	Y.P. 776 (265 ft.)	Y.P. 750A (221 ft.)	Y.P. 750B (221 ft.)	Y.P. 750C (221 ft.)	Y.P. 749–757 (221–236 ft.)
SiO_2	82.65	77.98	77.61	76.95	75.96	80.63	.....	1.35	55.78
Al_2O_3	10.68	11.60	11.74	12.18	12.21				
P_2O_5	0.07	0.13	0.04	0.07	0.11	9.98	.....	0.24	23.03
TiO_2	0.14	0.11	0.11	0.13	0.14				
Fe_2O_3	0.18	0.92	1.13	1.07	1.65	0.48 (Total Fe)	.....	0.37 (Total Fe)	5.05 (Total Fe)
FeO	0.08	0.10	0.21	0.26	0.36	.....	.....	.....	.....
MgO	0.05	0.06	0.14	0.12	0.01	Tr	.....	0.01	1.13
CaO	0.08	0.29	0.35	0.28	0.44	0.28	.....	0.02	1.08
MnO	Tr	Tr	0.00	0.01	0.01	.....	.....	.....	.....
Na_2O	0.22	2.56	2.85	2.98	2.90	2.72	3.58	.....	.....
K_2O	0.32	4.94	4.91	5.50	5.12	4.61	8.33	0.08	.....
BaO	0.00	Tr	n.d.	0.08	n.d.	0.03	.....	0.14	.....
Cr_2O_3	0.00	0.00	n.d.	n.d.	n.d.	.....	.....	.....	.....
ZrO_2	0.03	0.03	n.d.	0.03	n.d.	.....	.....	.....	.....
H_2O+	4.79	0.97	0.59	0.24	0.72	0.38 (By ignition)	.....	.....	5.08 (By ignition)
H_2O-	0.28	0.18	0.21	0.06	0.33				
CO_2	n.d.	0.00	0.00	0.04	0.01	.....	.....	.....	.....
Cl	n.d.	n.d.	n.d.	n.d.	n.d.	.....	.....	.....	.....
SO_3	0.98	.....	n.d.	.....	n.d.	.....	.....	.....	.....
S	.....	0.03	n.d.	0.41	n.d.	.....	.....	.....	.....
F	n.d.	n.d.	n.d.	0.00	n.d.	.....	.....	.....	.....
	100.55	99.90	99.89	100.41	99.97	99.11	11.91	2.21	96.45
Less oxygen for S	.....	.....	.....	0.15	.....	.....	.....	.....	.....
				100.26					

the feldspathic groundmass has been replaced by potash. Attack by CO_2 has subsequently leached alkalis from a small part of the feldspars, leaving a clay of the beidellite type. In Y.P. 669 still later action by acid waters has probably produced kaolinite. Y.P. 750A is phenocrysts of orthoclase and quartz picked out of crushed rock for the purpose of determining composition of the original feldspar

phenocrysts of the rhyolite. The quartz is extraneous, but it could not well be avoided. A little pyrite was included in the feldspars. Y.P. 750B is thoroughly metamorphosed groundmass, picked out of crushed specimen. Y.P. 750C is soluble material extracted by boiling the rock powder with dilute HCl, followed by $Na_2CO_3H_2O$. Y.P. 749–757 is beidellite from orthoclase casts. It doubtless contained a little quartz and feldspar.

WATER CIRCULATION IN GEYSER BASINS

Figure 4 shows that in the Upper Geyser Basin the steam pressure at any depth in the drill hole is not sufficient to overcome the weight of a column of water extending from the surface to that depth. In the upper part of the Norris hole the same condition was found, but in the last few feet a much higher pressure was met. This was doubtless due in part to steam from a lower region rising through a fissure, which provided relatively unrestricted escape.

If we assume that the fairly regular temperature gradient in the hole in the Upper Basin continues unchanged for some distance below the bottom of the hole, and plot the corresponding steam pressures, we find that at about 500 feet the graph so constructed intersects the graph of hydrostatic pressures of the water column. From this intersection downward the two graphs rapidly diverge. At 600 feet steam pressure exceeds hydrostatic pressure by more than 50 pounds per square inch; at 700 feet the excess is more than 130 pounds, and the absolute value of steam pressure is more than 400 pounds per square inch. These figures should not be regarded as giving more than an approximation, for doubtless there are irregularities, but they indicate the great pressures that probably exist. The free escape of the steam from underground is prevented by the impervious character of the overlying rock and the resistance to expulsion of the water in the interstices, but if the hydrostatic head were no greater than that of a water column extending downward from the surface at the drill site, the condition could be only a temporary one. At any depth greater than 500 feet the steam pressure would be greater than the hydrostatic pressure. The water in the interstices of the rock would be gradually expelled, and magmatic vapors would work their way to the surface. In order to maintain the con-

dition that exists, it seems necessary that meteoric water from a somewhat distant source, under greater hydrostatic head than the steam pressure, should continually replenish the supply.

At Norris, at a distance of 2 miles on the south and 4 miles on the east and northeast, large areas of the rhyolite plateau have an elevation 600 feet higher than the drill site. We do not believe these distances too great for a small quantity of water to find its way through the rocks. The waters of many ordinary springs are believed to travel greater distances. The chief essential difference in principle between the circulation of ordinary springs and that of the geyser basins is that in ordinary springs the depth to which waters penetrate before rising again to the surface is limited principally by friction in passing through the rocks, whereas in the circulation that enters geyser basins there is in addition the limitation that the water cannot penetrate downward to a greater depth than that at which the steam pressure equals the hydrostatic head.

In the Upper Geyser Basin areas of high plateau, larger than those in the vicinity of Norris, with a somewhat greater differential elevation, lie a shorter distance away. Consequently, the flow of water in deep circulation is probably greater. At Norris the body of underground water is more nearly stagnant and its depth of penetration is less.

Under the assumption previously made, that the temperature gradient in the Upper Basin continues uniform to a greater depth, we find that the pressure of saturated steam at 800–900 feet is sufficient to balance the hydrostatic head of a column of water coming from a source 600 feet higher than the site of the drill hole. This probably represents a maximum figure for the depth of penetration of meteoric water. The temperature here is more than 300°, and steam pressure increases so rapidly with increase of temperature that 100 feet additional hydrostatic head would not enable the water to penetrate much deeper. Furthermore, the assumption regarding temperature gradient probably errs in being too small at the greater depth, for the flow of water that carries the heat away naturally diminishes with depth.

At both basins one must travel several miles before reaching a point on the surface where the elevation is as low as the bottoms of

the respective holes. Probably any meteoric water that enters the basin areas to a depth as great as the bottom of either hole has little chance of traveling horizontally to such points of egress. Most of it must be deflected upward by the heat and pressure imparted to it, and rise to the surface within the basins. The surface outflow from each basin is therefore a measure of the amount of water that comes in from the adjacent plateau, and we find, as we should expect, that the outflow from Norris Basin is much less than that from the Upper Basin. This has been established by stream measurements made by Allen. He has attached great importance to such factors as those discussed here, and he was the first to recognize their fundamental significance in Yellowstone phenomena.

We believe that the difference in the amount of inflow and the depth reached in the respective basins governs the character of metamorphic processes. In the Upper Basin a large inflow descends to a great depth. The CO_2 that it gains from magmatic sources has opportunity to act upon a great thickness of rock before the waters rise to the level of the drill hole, and therefore they have acquired by this time their full complement of alkalis and silica. Their further action is to substitute potash for soda and to deposit silica. They arrive at the surface in an alkaline condition.

In Norris Basin the amount of water and the depth are less. The action of CO_2 upon the rock is not carried to completion before the waters reach the zone penetrated by the drill hole; therefore they attack the feldspars within this zone and leave a clay. They arrive close to the surface with only a small content of alkali bicarbonate. Within this uppermost zone they mingle with shallow surface water, in which H_2S becomes oxidized to H_2SO_4, and their slight alkalinity is more than neutralized. According to the interpretation given, the attack of CO_2 upon the rocks, by which beidellite (or some other alkali-poor and alumina-rich mineral) is formed, and alkaline bicarbonates taken into solution, began where the rising gases met the ground water, and the zone of attack is gradually rising from lower to higher levels rather than descending from higher to lower. This is the conclusion to which petrographic and chemical studies of the cores from the two drill holes seem to point.

In the Upper Basin the issuing waters have a high content of alkali bicarbonate and little free CO_2. Since there is no evidence in the cores from this hole that the feldspathic material is being altered to clay, Allen has drawn the logical conclusion that the zone of attack must now lie below the region penetrated by the drill hole. The petrographic evidence, however, seems to go farther and indicate that the zone of attack must always have been below this level, for if these upper rocks had been subjected to attack at some previous time, while the zone was descending from a higher level to its present position, evidence should have been left. Even if the products formed had been carried away in suspension, empty casts of feldspar phenocrysts should be visible (as in the Norris cores), and none are seen.

In the Norris cores it appears that the alteration of feldspars to beidellite is now in operation. It has followed rather than preceded base exchange in the feldspars. It is inferred that the same sequence will ultimately be established in the rocks of the Upper Basin by the rise of the zone of CO_2 attack from lower levels. The appearance of clay at higher levels at Norris earlier than in the Upper Basin is attributed to the shallower depth of circulation of meteoric waters at Norris, and consequently an earlier exhaustion (or partial exhaustion) of material readily acted upon by CO_2 in the regions below. This lack of easily attacked material at lower levels is believed to be attested by the deficiency of alkaline bicarbonates and excess of free CO_2 in the issuing waters of this basin.

Thus a more advanced stage of alteration appears to have been reached now at Norris than in the Upper Basin, but the stage of base exchange was less complete at Norris and was probably passed through more quickly.[24]

ESSENTIAL FEATURES OF GEYSER ACTION

The drilling of the holes in the geyser basins has given much information on the structure of the rocks and the thermal state of the

[24] For an interpretation of the significance of beidellite, differing in some respects from that given here, see E. T. Allen and Arthur L. Day, *Hot Springs of the Yellowstone National Park*, p. 490.

water that fills its pores, crevices, and cavities. It supplies a more complete and definite picture of underground conditions than has been available heretofore.

The manner in which pools of hot water in the neighborhood of the drill site in the Upper Geyser Basin were affected when large volumes of steam were allowed to be discharged from the hole; and the breaking-out of vents around the Norris site when steam under high pressure was tapped indicate a complex system of passages, some of them freely intercommunicating and others restricted.

The popular conception of a geyser tube apparently pictures it as a fairly regular orifice of simple structure. Textbook discussions of geyser action are likely to be defective in not giving an adequate explanation of what occurs after an eruption is over, or by what means the superheated steam that issues forcibly from the rock surrounding the empty conduit is restrained.

The geyser system must be a complicated affair, and one of the essentials is that is must be so constructed that at the end of an eruption a body of relatively cool water (at less than boiling temperature) must be available in sufficient quantity to quench the steam that leaks into the cavities.

The system suggested by these requirements and indicated by the drilling is one in which an irregular main conduit extends from the surface to a moderate depth, and is connected more or less freely with ramifying fissures, channels, and cavities extending upward, downward, and horizontally. Some portion of the main channel (not necessarily its lowest point) is more directly connected with the heat supply than subsidiary channels, and therefore the water in it becomes heated more rapidly to the temperature at which its vapor pressure is able to lift the load.

When the water becomes hot enough to develop steam bubbles and lift the overlying column, causing an overflow at the surface, or a rise of water in the funnel-like opening, pressure is slightly relieved in the tube. There are usually a number of these preliminary symptoms before the eruption gets under way. We believe that when pressure is relieved in the manner suggested, slightly cooler water in connecting channels moves toward the region of lessened pressure, and causes condensation, and thus produces a few moments' delay.

There may be a number of these surges, but finally the eruption occurs. With the ejection of the column of water and the consequent great relief of pressure, quantities of water from connecting channels pour into the fissure. This may be water cool enough to condense the steam, or it may be heated water that is ejected immediately. The jet may die down entirely for a few moments, and then start again in full vigor. In the Grand Geyser the writer has seen twenty-one separate discharges of this kind in the course of an eruption.

When the eruption definitely ceases the water that enters the cavities, either as a small trickle or in large volume, must be below the boiling-point for atmospheric pressure.